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PAPERS IN PRESS FOR FORTHCOMING ISSUES

Belt Series of the North: Stratigraphy, Sediments, Fossils. By C. L. Fenton and M. A. Fenton

Stratigraphy of the Rio Querecual Section of Northeastern Venezuela. By Hollis D. Hedberg

Stratigraphy and Paleontology of the Upper Cretaceous Beds Along the Eastern Side of Laguna de Mayran, Coahuila, Mexico. By Ralph W. Imlay

Silurian Strata of Notre Dame Bay and Exploits Valley, Newfoundland. By W. H. Twenhofel and R. R. Shrock

Silurian Strata of White Bay, Newfoundland. By George R. Heyl

Glacial Geology of Mauna Kea, Hawaii. By H. E. Gregory and C. K. Wentworth

Geology of Sierra de la Peña. By Theodore S. Jones

Laramide Igneous Sequence and Differentiation in the Front Range, Colorado. By T. S. Lovering and E. N. Goodard

Vein-filling at Nevada City, California. By W. D. Johnston, Jr.

Correlation of the Tyler and the Copps Formations of the Gogebic Iron District. By Gordon I. Atwater

Earthquake Investigations in the Western Mountain Region. By N. H. Heck

FIGURE 1. DARK INCLUSIONS IN BOULDER BATHOLITH
More abundant than average but otherwise typical of inclusions in batholiths.

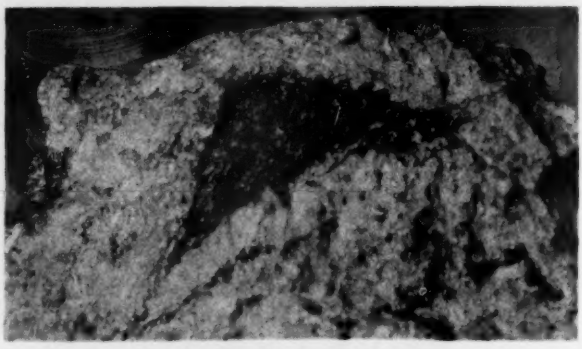

FIGURE 2. VERMILION GRANITE
Mica schist inclusion partly broken down to xenocrysts and strewn into granite.

FIGURE 3. BORDER OF GRANITE AT SEA POINT, SOUTH AFRICA
Inclusions of slate, and large feldspar phenocrysts.

INCLUSIONS IN GRANITES

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
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CRITERIA OF ORIGIN OF INCLUSIONS IN PLUTONIC ROCKS

BY FRANK F. GROUT

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INTRODUCTION

In 1922, N. L. Bowen presented his carefully considered opinions and supporting data on the behavior of inclusions in magmas. That paper was one of the finest contributions of this able and versatile petrologist. Its ideas have been widely accepted, considerably discussed, and only slightly, if at all, modified. Nevertheless, field geologists who find inclusions in magmatic rocks do not always find convincing evidence as to origins, and even those who accept Bowen's ideas may, with lack of data in a specific instance, ignore those ideas or assume that the evidence favors one or another origin. The evidences used need critical discussion. This study offers a summary of criteria and gives some new data for several districts. Emphasis is given to criteria used since Bowen's paper was issued. Most of the problems lie in the dark spots, "black knots," of granitic or dioritic rocks, but some attention is given to others also.

Bowen's final summary calls attention to the reaction of inclusions and magma to form minerals like those growing in the magma. He says:

> "It is this action of magma on inclusions that makes particularly difficult the problem of distinguishing xenolith from autolith, i. e., accidental inclusion from cognate inclusion."[1]

It must be borne in mind that criteria are not all of equal significance, and that some may be of service in a restricted district but not universally, that many are only suggestive and not at all conclusive except when a whole series of suggestions point the same way, and that when several criteria give conflicting suggestions the value of any one specific criterion is greatly reduced. It can hardly be too strongly emphasized that no single criterion is commonly sufficient to prove an origin. If one insisted on using only those criteria to which there are no exceptions,

[1] N. L. Bowen: *The behavior of inclusions in igneous magmas*, Jour. Geol., vol. 30 (1922) p. 539.

the list would be brief, and few occurrences would yield conclusive data. Nevertheless, when there is a noteworthy convergence of evidence based on several features the indicated origin may commonly be accepted.

The writer's field and laboratory studies of the problem include several districts. These studies and some of the analyses have been made possible by support from the Minnesota Geological Survey, and the research funds of the Graduate School of the University of Minnesota to whom grateful acknowledgement is here offered.

WORKS DISCUSSED

Even before 1880, many papers had been written concerning inclusions in plutonic rocks. A selected list is here given of those which have discussed the origins of such bodies or suggested criteria for their determination.

1880. PHILLIPS, J. A.: *On concretionary patches and fragments of other rocks in granite,* Geol. Soc. London, Quart. Jour., vol. 36, p. 1-22.

1888. WILLIAMS, G. H.: *The contact-metamorphism produced in the adjoining mica schists and limestones by the massive rocks of the "Cortlandt series" near Peekskill, N. Y.,* Am. Jour. Sci., 3rd series, vol. 36, p. 254-269.

TEALL, J. J. H.: *British petrography,* p. 311.

1890. LACROIX, A.: [A series of papers on inclusions in volcanic rocks]. Soc. Géol. France, Bull., vol. 18, p. 845-880.

BROEGGER, W. C.: *Die Mineralien der Syenitpegmatitgänge der Südnorwegischen Augit- und Nephelinsyenite,* Zeit. für Krist. und Min., vol. 16, p. 105-106.

1893. ZIRKEL, F.: *Lehrbuch der Petrographie,* I, p. 504, 712, 787-796.

1894. TURNER, H. W.: *The rocks of the Sierra Nevada,* U. S. Geol. Surv., 14th Ann. Rept., pt. 2, p. 475-481, pl. 54.

VON CHRUSTSCHOFF, K., *Über holokrystalline makrovariolitische Gesteine,* Acad. Sci., St. Petersburg, Mem., ser. 7, vol. 42, no. 3, p. 230-1.

1894-1895. GRIMSLEY, G. P.: *The granites of Cecil County, Maryland,* Cincinnati Soc. Nat. Hist., Jour. vol. 17, p. 59-67, 78-90.

1898. LACROIX, A.: *Les modifications endomorphes du gabbro du Pallet (Loire-Inférieure),* Acad. Sci. Paris, C. R. 126, p. 1038-1040.

1899. CROSS, WHITMAN, and PURINGTON, C. W.: *Telluride, Colorado,* in *Geologic Atlas of the United States,* U. S. Geol. Surv., folio 57, p. 7.

1900. HOLLAND, T. H.: *The charnockite series, a group of Archæan hypersthenic rocks in peninsular India,* Geol. Surv. India, Mem. 28, pt. 2, p. 215-218, 235-248.

1903. (a) DALY, R. A.: *The mechanics of igneous intrusion,* Am. Jour. Sci., 4th ser., vol. 16, p. 108.

(b) ——: *The geology of Ascutney Mountain, Vermont,* U. S. Geol. Surv., Bull. 209, p. 43-84.

1904-1906. KNOPF, A., and THELEN, P.: *Sketch of the geology of Mineral King, California,* Univ. Calif., Publ., Bull. Dept. Geol., vol. 4, no. 12, p. 227-262.

1906. GILBERT, G. K.: *Gravitational assemblage in granite,* Geol. Soc. Am. Bull., vol. 17, p. 321-328.

1907. Rosenbusch, H.: *Mikroskopische Physiographie der Mineralien und Gesteine,* vol. 2, pt. 1, p. 92.

1908. Wright, F. E., and Wright, C. W.: *The Ketchikan and Wrangell mining districts, Alaska,* U. S. Geol. Surv., Bull. 347, p. 63.

Bastin, E. S.: *Rockland, Maine,* in *Geologic Atlas of the United States,* U. S. Geol. Surv., folio 158, p. 7.

1909. Harker, A.: *The natural history of igneous rocks,* p. 322, 348-349.

Iddings, J. P.: *Igneous rocks,* vol. I, p. 253-254.

1910. Adams, F. D., and Barlow, A. E.: *Geology of the Haliburton and Bancroft areas, Province of Ontario,* Geol. Surv., Canada, Mem. 6, p. 62-87, 120-129.

Merrill, G. P.: *Stone for building and decoration,* 3rd edition, p. 65-72, 537.

1911. Rogers, G. S.: *Geology of the Cortlandt series and its emery deposits,* New York Acad. Sci., Ann. 21, p. 52-54, 81.

Hall, A. L., and Gardthausen, C.: *Note on some remarkable zenoliths of altered shale from the norite of Potgietersrust and Mapoch's country,* Geol. Soc. South Africa, Tr., vol. 14, p. 74-78.

Knopf, A.: *Geology of the Berners Bay region, Alaska,* U. S. Geol. Surv., Bull. 446, p. 23.

Loewinson-Lessing, F.: *The fundamental problems of petrogenesis, or the origin of the igneous rocks,* Geol. Mag., vol. 48, p. 289-297.

1912. Goldschmidt, V. M.: *Die Kontaktmetamorphose im Kristianiagebiet,* Videnskabs-selskab. Skrifter Math.-Naturv., Kl. 199, no. 1, p. 105-116, 292-303.

Loughlin, G. F.: *The gabbros and associated rocks at Preston, Connecticut,* U. S. Geol. Surv., Bull. 492, p. 39-40.

Reid, C., et al: *The geology of Dartmoor,* Geol. Surv. England and Wales, explanation of sheet 338, p. 41.

Weed, W. H.: *Geology and ore deposits of the Butte district, Montana,* U. S. Geol. Surv., Prof. Paper 74, p. 32, 38.

1912-1913. Daly, R. A.: *Geology of the North American Cordillera at the Forty-ninth Parallel,* Canada Dept. Interior, Rept. Chief Astronomer, p. 243-244, 298-303, 349, 457, 461, 526-539, 735.

1913. de Lapparent, J.: *Les enclaves du granite du Mont Haya en pays basque,* Soc. Français Minéralogie, Bull., vol. 36, p. 163-169.

Taber, S.: *Geology of the gold belt in the James River basin, Virginia,* Va. Geol. Surv., Bull. 7, p. 103-105.

1914. Wagner, P.: *The geology of a portion of the Belingwe district of Southern Rhodesia,* Geol. Soc. South Africa, Tr., vol. 17, p. 46.

Winchell, A. N.: *Mining districts of the Dillon quadrangle, Montana, and adjacent areas,* U. S. Geol. Surv., Bull. 574, p. 148-151.

Allan, J. A.: *Geology of Field map-area, B. C. and Alberta,* Geol. Surv. Canada, Mem. 55, p. 189-190.

Eskola, P.: *On the petrology of the Orijärvi region in southwestern Finland,* Comm. Geol. Finlande, Bull. no. 40, p. 64-68.

Miller, W. J.: *Magmatic differentiation and assimilation in the Adirondack region,* Geol. Soc. Am., Bull., vol. 25, p. 254-264.

1915. Powers, Sidney: *The origin of the inclusions in dikes,* Jour. Geol., vol. 23, p. 2, 180-181.

1915-1916. Billingsley, Paul: *The Boulder batholith of Montana,* Am. Inst. Min. Engs., Tr., vol. 51, p. 41-42.

1916. Tyrrell, G. W.: *A contribution to the petrography of Benguella,* Roy. Soc., Edinburgh, Tr., vol. 51, pt. 3, p. 543-545.

1917. Collins, W. H.: *Onaping map-area,* Geol. Surv. Canada, Mem. 95, p. 54-55.
1917-1918. Billingsley, Paul, and Grimes, J. A.: *Ore deposits of the Boulder batholith of Montana,* Am. Inst. Min. Engs., Tr., vol. 58, p. 289.
1918. Wilson, M. E.: *Timiskaming County, Quebec,* Geol. Surv. Canada, Mem. 103, p. 130.
Bowles, O.: *The structural and ornamental stones of Minnesota,* U. S. Geol. Surv., Bull. 663, p. 49-115.
Knopf, A.: *A geologic reconnaissance of the Inyo Range, and the eastern slope of the southern Sierra Nevada, California,* U. S. Geol. Surv., Prof. Paper 110, p. 64-65.
1921. Tilley, C. E.: *The granite-gneisses of southern Eyre Peninsula (South Australia) and their associated amphibolites,* Geol. Soc. London, Quart. Jour., vol. 77, p. 91-93.
Goldschmidt, V. M.: *Geologisch petrographischen Studien im Hochgebirge des südlichen Norwegens, V. Die Injectionsmetamorphism im Stavanger-Gebiete,* Skriften Vidensk. I. Kristiania, 1920, 2 Band, no. 10, p. 49-114.
Clapp, C. H.: *Geology of the igneous rocks of Essex County, Massachusetts,* U. S. Geol. Surv., Bull. 704, p. 46, 114-115.
1922. Thomas, H. H.: *On certain xenolithic Tertiary minor intrusions in the Island of Mull (Argyllshire),* Geol. Soc. London, Quart. Jour., vol. 78, p. 229-260.
Bowen, N. L.: *The behavior of inclusions in igneous magmas;* Jour. Geol., vol. 30 p. 513-570.
1923. Beger, P. J.: in P. Niggli's *Gesteins- und Mineralprovinzen,* p. 461-467.
Read, H. H.: *The petrology of the Arnage district in Aberdeenshire: a study of assimilation,* Geol. Soc. London, Quart. Jour., vol. 79, p. 473-479.
Dale, T. N.: *The commercial granites of New England,* U. S. Geol. Surv., Bull. 738, p. 58-65.
Sederholm, J. J.: *On migmatites and associated pre-Cambrian rocks of southwestern Finland,* Comm. Geol. Finlande, Bull., no. 58 p. 113-128.
1924. Bailey, E. B., et al: *Tertiary and post-Tertiary geology of Mull, Loch Aline, and Oban,* Geol. Surv. Scotland, Mem., p. 34, 212-369.
Read, H. H.: *On certain xenoliths associated with the contaminated rocks of Huntly mass, Aberdeenshire,* Geol. Mag., vol. 61, p. 435-440.
1925. Allison, I. S.: *The Giants Range batholith of Minnesota,* Jour. Geol., vol. 33, p. 508.
Dixey, F.: *The geology of Sierra Leone,* Geol. Soc. London, Quart. Jour., vol. 81, p. 205.
Ries, H., and Watson, T. L.: *Engineering geology,* 3rd edition, p. 480.
Martin, Laura H.: *The geology of the Stonington region, Connecticut,* Conn. Geol. Nat. Hist. Surv., Bull. 33, p. 27-29.
1926. Grout, F. F.: *The geology and magnetite deposits of northern St. Louis County, Minnesota,* Minn. Geol. Surv., Bull. 21, p. 29-52.
Smyth, C. H., and Buddington, A. F.: *Geology of the Lake Bonaparte quadangle,* N. Y. State Mus., Bull. 269, p. 79-90, pls. 23, 24.
Pirsson, L. V., (A. Knopf): *Rocks and rock minerals,* 2nd edition, p. 168-169.
Read, H. H., et al: *The geology of Strath Oykell and lower Loch Shin,* Geol. Surv. Scotland, Mem., p. 55-103.
Emmons, R. C.: *Concerning inclusions in igneous magmas,* Jour. Geol., vol. 34, p. 425-427.
Sederholm, J. J.: *On migmatites and associated pre-Cambrian rocks of southwestern Finland,* II, Comm. Geol. Finlande, Bull. no. 77, p. 18-22, 102-103.

1927. GHOSH, P. K.: *Petrology of Bodmin Moor granite (eastern part), Cornwall,* Mineral. Mag., vol. 21, p. 302-308.

1928. PABST, A.: *Observations on inclusions in the granitic rocks of the Sierra Nevada,* Univ. Calif., Publ., Bull. Dept. Geol. Sci., vol. 17, p. 325-386.

GRANTHAM, D. R.: *The petrology of the Shap granite,* Geol. Assn., Pr., vol. 39, p. 303-327.

STANSFIELD, J.: *Assimilation and petrogenesis,* p. 27-40.

DALY, R. A.: *Bushveld igneous complex of the Transvaal,* Geol. Soc. Am., Bull., vol. 39, p. 746-750.

TYRRELL, G. W.: *The geology of Arran,* Geol. Surv. Scotland, Mem., p. 124-200.

1929. SANDERS, C. W.: *A composite stock at Snowbank Lake in northwestern Minnesota,* Jour. Geol., vol. 37, p. 135-149.

BUDDINGTON, A. F., and CHAPIN, T.: *Geology and mineral deposits of southeastern Alaska,* U. S. Geol. Surv., Bull. 800, p. 178.

GROUT, F. F.: *The Saganaga granite of Minnesota-Ontario,* Jour. Geol., vol. 37, p. 580-587.

GOODSPEED, G. E.: *The mode of origin of a reaction porphyry dike at Cornucopia, Oregon,* Jour. Geol., vol. 37, p. 161-175.

1930. GOODSPEED, G. E.: *Some effects of the recrystallization of xenoliths at Cornucopia, Oregon,* Am. Jour. Sci., 5th ser., vol. 20, p. 145-150.

RICHEY, J. E., and THOMAS, H. H.: *The geology of Ardnamurchan, northwest Mull and Coll,* Geol. Surv. Scotland, Mem., p. 100-102.

JONES, W. A.: *A study of certain xenoliths occurring in gabbro at Sudbury, Ontario,* Univ. Toronto Studies, geol. ser., no. 29, p. 61-73.

VOGT, J. H. L.: *The physical chemistry of the magmatic differentiation of igneous rocks,* III, 2nd half, Skr. Norske Videnskaps-Akad., Oslo, 1930, Math-Naturv. Klasse 1930, no. 3, p. 68.

SUGI, K.: *On the granitic rocks of Tsukuba district,* Japanese Jour. Geol. Geog., Tr., vol. 8, p. 43-52, 92.

MACGREGOR, A. M.: *The geology of the country between Gatooma and Battlefields,* Southern Rhodesia Geol. Surv., Bull. 17, p. 49-50.

1931. CLOOS, E.: *Der Sierra Nevada-Pluton,* Geol. Rundschau, vol. 22,, p. 372-384.

MILLER, W. J.: *Geologic sections across the southern Sierra Nevada of California,* Univ. Calif., Publ., Bull. Dept. Geol. Sci., vol. 20, p. 331-360, pl. 58.

NOCKOLDS, J. R.: *The Dhoon (Isle of Man) granite: a study in contamination,* Mineral. Mag., vol. 22, p. 504-505.

MACGREGOR, A. G.: *Scottish pyroxene-granulite hornfelses and Odenwald beerbachites,* Geol. Mag., vol. 68, p. 506-521.

LAHEE, F. H.: *Field geology,* 3rd edition, p. 141.

JEREMINE, E.: *Quelques nouvelles données chimiques et minéralogiques sur le granite et les roches métamorphiques de Flamanville,* Soc. Français Minéralogie, Bull., vol. 54, p. 25-46.

WELLS, A. K., and WOOLDRIDGE, S. W.: *The rock groups of Jersey,* Geol. Assn., Pr., vol. 42, p. 185-205.

1932. FITCH, A. A.: *Contact metamorphism in southeastern Dartmoor,* Geol. Soc. London, Quart. Jour., vol. 88, p. 582.

VON GOTTFRIED, C.: *Ueber endogene basische Einschlüsse in sauren Tiefengesteinen,* I, *Adamello, Chemie der Erde,* vol. 7, p. 586-593.

NOCKOLDS, S. R.: *The contaminated granite of Bibette Head, Alderney,* Geol. Mag., vol. 69, p. 438-452.

RICHEY, J. E., and THOMAS, H. H.: *The Tertiary ring complex of Slieve Gullion . . .,* Geol. Soc. London, Quart. Jour., vol. 88, p. 776-849.

TIDMARSH, W. G.: *The Permian lavas of Devon,* Geol. Soc. London, Quart. Jour., vol. 88, p. 712-775.

DUTOIT, A. L.: *How the kimberlite pipes serve to reveal the unknown,* Econ. Geol., vol. 27, p. 206-209.

JOHANNSEN, A.: *A descriptive petrography of the igneous rocks,* II, p. 248-249, 259, 354.

GARDINER, C. I., and REYNOLDS, S. H.: *The Loch Doon (granite) area, Galloway,* Geol. Soc. London, Quart. Jour., vol. 88, p. 1-34.

THOMAS, H. H., and SMITH, W. C.: *Xenoliths of igneous origin in the Trégastel-Ploumanach'h granite, Cotes du nord, France,* Geol. Soc. London, Quart. Jour., vol. 88, p. 276-296.

BRAMMALL, A., and HARWOOD, H. F.: *The Dartmoor granites: their genetic relationships,* Geol. Soc. London, Quart. Jour., vol. 88, p. 202-225.

WILLIAMS, A. F.: *The genesis of the diamond,* I, p. 196-352; II, p. 353-402.

1933. JOPLIN, G. A.: *Petrology of the Hartley district,* II, Linnean Society, New South Wales, Pr., vol. 58, p. 132-146.

CLOOS, E.: *Structure of the "Ellicott City granite", Maryland,* Nat. Acad. Sci., Pr., vol. 19, p. 131-132.

BRAMMALL, A.: *Syntexis and differentiation,* Geol. Mag., vol. 70, p. 98-100.

GROUT, F. F.: *Contact metamorphism of the slates of Minnesota by granite and by gabbro magmas,* Geol. Soc. Am., Bull., vol. 44, p. 1008-1033.

TANSLEY, W., SCHAFER, P. A., and HART, L. H.: *A geological reconnaissance of the Tobacco Root Mountains, Madison County, Montana,* Mont. Bur. Mines and Geol., Mem. 9, p. 15.

NOCKOLDS, S. R.: *Some theoretical aspects of contamination in acid magmas,* Jour. Geol., vol. 41, p. 561-589.

1934. SEDERHOLM, J. J.: *On migmatites and associated pre-Cambrian rocks of southwestern Finland,* III, Geol. Komm. Finland, Bull. 107, p. 37-41.

BOWLES, O.: *The stone industries,* 1st ed., p. 110-111.

NOCKOLDS, S. R.: *The contaminated tonalites of Loch Awe, Argyll,* Geol. Soc. London, Quart. Jour., vol. 90, p. 302-321.

REYNOLDS, D. L.: *The eastern end of the Newry igneous complex,* Geol. Soc. London, Quart. Jour., vol. 90, p. 591, 625.

MILLER, W. J.: *Geology of the western San Gabriel Mountains of California,* Univ. Calif. at Los Angeles, Publ. in Math. and Physical Sci., vol. 1, no. 1, p. 1-114.

BALK, ROBERT, and GROUT, F. F.: *Structural study of the Snowbank stock,* Geol. Soc. Am., Bull., vol. 45, p. 621-636, figs. 4, 8.

GHOSH, P. K.: *The Carnmenellis granite,* Geol. Soc. London, Quart. Jour., vol. 90, p. 240-276.

1935. SCHOKLITSCH, KARL: *Pyrometamorphose an Einschlüssen in Eruptivgesteine der Alpen-Ostrand,* Tschermaks Min. Pet. Mitth., vol. 46, p. 127-152.

WALKER, F., and DAVIDSON, C. F.: *Marginal and contact phenomena of the Dorback granite,* Geol. Mag., vol. 72, p. 52, 61-62.

NOLAN, T. B.: *The Gold Hill mining district, Utah,* U. S. Geol. Surv., Prof. Paper 177, p. 45.

1935. MAYO, E. B.: *Some instrusions and their wall rocks in the Sierra Nevada,* Jour. Geol., vol. 43, p. 682.

HURLBUT, C. S., JR.: *Dark inclusions in a tonalite of southern California,* Am. Mineral., vol. 20, p. 609-630.

JOPLIN, G. A.: *A note on the origin of basic xenoliths in plutonic rocks, with special reference to their grain size,* Geol. Mag., vol. 72, p. 227-234.

CALLAGHAN, EUGENE: *Pre-granodioritic dikes in granodiorite, Paradise Range, Nevada,* Am. Geophysical Union, Tr., 16th annual meeting, I, p. 302-307.

THOMSON, J. E.: *On the origin of the syenites of Coldwell, Ontario,* Am. Geophysical Union, Tr., 16th annual meeting, I, p. 350-363.

WILLIAMSON, W. O.: *The composite gneiss and contaminated granodiorite of Glen Shee, Perthshire,* Geol. Soc. London, Quart. Jour., vol. 91, p. 382-422.

COLLINS, W. H.: *Derivations of granitic rocks,* Report of the 16th Intern. Geol. Cong., preprint, p. 271-282.

READ, H. H.: *The gabbros and associated xenolithic complexes of the Haddo House district, Aberdeenshire,* Geol. Soc. London, Quart. Jour., vol. 91, p. 592-638.

1935-1937. NOCKOLDS, S. R.: *Contributions to the petrology of Barnavave, Carlingford, I. F. S.—I. The junction hybrids,* Geol. Mag., vol. 72, 1935, p. 289-315; II, vol. 74, 1937, p. 125-132.

1936. IWAO, S.: *On some basic inclusions in granite of Kuga district, Nagato; a study in contamination,* Japanese Jour. Geol. Geog., vol 13, p. 155-162.

KENNEDY, W. Q., and READ, H. H.: *The differentiated dike of Newmains, Dumfriesshire, and its contact and contamination phenomena,* Geol. Soc. London, Quart. Jour., vol. 92, p. 116-144.

HEYL, G. R.: *Geology and mineral deposits of the Bay of Exploits area,* Newfoundland Dep. Nat. Res., Geol. Sect., Bull. 3, p. 26-27.

SMITH, H. G.: *The South Hill lamprophyre, Jersey,* Geol. Mag., vol. 73, p. 87-91.

MACGREGOR, MALCOLM: *A xenolithic pegmatite in the Dalbeattic "Granite",* Geol. Mag., vol. 73, p. 171-185.

TERMS USED

So many different terms have been applied to inclusions in granites and related rocks that a list of terms is necessary before the origin of the inclusions can be satisfactorily discussed. Many have been used in a descriptive sense, but certain authors may have had in mind a particular origin when using a particular term. With some exceptions, common usage is indicated by the headings in the following list:

GENERAL TERMS NOT IMPLYING ORIGIN: Inclusions, basic inclusions, enclosures, inclosures, knots, clots, dark spots, niggerheads, bunches, enclaves, spots, nests of mineral.

IMPLYING A SOURCE OUTSIDE THE INTRUSIVE: Xenoliths (and xenocrysts), foreign inclusions, exogenous inclusions, accidental inclusions, enclaves énallogènes.

IMPLYING A SOURCE IN THE IGNEOUS ROCK MASS ITSELF: Indigenous spots, endogenous inclusions, schlieren, cognate inclusions, consanguineous inclusions, enclaves homeogènes.

(A) Segregations, basic segregations, secretions, lateral secretions, concretions, nodules, autoliths.

(B) Cognate fragments, endogenous fragments, protoclastic structures; even cognate xenoliths,[3] but that seems to contradict itself. (Autoliths, placed here by some writers, are best restricted to segregations.)

[3] The writer has used the expression [*Petrography and petrology* (1932) p. 36] but now considers it an erroneous usage.

IMPLYING A MIXED ORIGIN: Reaction inclusion, hybrid inclusion, migmatite inclusion.

Many of these terms have been defined in different ways, and there is much confusion. Only a few are needed, and possibly some agreement can be reached on those. Cognate materials are those formed inside the magma chamber. Xenoliths are rocks from outside, included in the magma. With these meanings, a "cognate xenolith" is an impossibility; the term has been used, probably unwisely, for fragments of early border facies. A segregation should imply that a mass grew, rather than was broken or corroded, into its present form.

Holmes [3] would like to see the term inclusion restricted to material inside mineral grains, leaving enclosures for the rocks here discussed; but this is not general usage. Lacroix proposed "enclaves homeogènes" with accurate definition, but, if translated "homogeneous inclusion," most geologists would consider that this implied uniformity in its own parts, not in its relation to its matrix. Holland (1900) proposed the term autolith for well-defined segregative schlieren, or segregations, to contrast with xenoliths. Mayo (1935) includes cognate fragments as autoliths, but this is not the original usage and probably is not wise. Holland's evidence, however, that certain dark inclusions in a granitoid rock were segregated, is weak. Probably the term, like the term segregation, is best reserved for objects concerning which the evidence is strong.[4]

There are many references to segregations, accompanied by descriptions, showing that the objects are fragments of early facies. The history of a segregated lump and that of a broken piece of an early border facies are so different that two different terms should be used if the two kinds can be distinguished in practice.

Up to about 1900, inclusions were partly comprehended in the word schlieren, signifying various irregularities in an igneous mass, which are genetically related; the term has become current, however, for streaky irregularities whether they are related or not.

Many other correlations or restrictions have been suggested, but none can be selected as general usage.[5] There are, of course, examples of careless statement or loose usage which add to the confusion.[6]

[3] Arthur Holmes: *The nomenclature of petrology*, 2d ed. (1928).

[4] Pabst (1928) uses "autolith" for all inclusions which "have not been shown to be" of other origin. This tacitly assumes an origin wherever the evidence against it has not been found and is a confusing usage.

[5] Ries and Watson (1925, p. 480) say that knots are segregations but that inclusions are different. Lahee (1931, p. 142) says segregations are to be distinguished from inclusions; also that cognate inclusions grade into segregations. Bowles (1934, p. 110) makes a similar point.

[6] E. H. L. Schwarz [*The Sea-Point granite-slate contact*, Geol. Soc. South Africa, Tr., vol. 16 (1913) p. 38] writes that "slate in the granite . . . would become a basic segregation." Laura H. Martin (1925, p. 29) says "inclusions . . . are garnetiferous segregations." G. A. Joplin (1933, p. 125) writes of "hybrids of the nature of segregations."

ANALYSIS OF THE PROBLEM OF ORIGINS

The purpose here is to discuss only slightly those inclusions that are recognized almost at once as derived from known wall rocks of the intrusive. Such rocks as recognizable shale, slate, mica schist, chert, marble, and quartzite are classed at once as xenoliths and need little study here.

The problems arise chiefly in rocks which have suffered contact metamorphism and exchanged constituents with the magma until they resemble igneous rocks (Pl. 1, fig. 1). A hornfels, with granoblastic, porphyroblastic, or poikiloblastic texture, and with the minerals of an igneous rock, may be indistinguishable in hand specimens from an igneous rock. Many detailed features should be studied in such igneous-looking inclusions to see if the origin can be determined. In this determination, distinctions should be made systematically in the following order:

1. Is the inclusion of sedimentary or of igneous origin?
2. If igneous, is it from outside (a xenolith), or is it of cognate origin?
3. If cognate, is it a fragment of an early facies or dike, or is it a segregation?

If the criteria of sedimentary origin are clear, the problem is solved. If the inclusion is of igneous origin, the work must be carried to the next alternative. If it is a xenolith, this settles the problem; but if cognate, the third group of alternatives should be worked through.

In several kinds of inclusions the recognition of the origin is complicated by reaction or hybridization. Criteria for these processes were once discussed by Richey and Thomas (1930), and these will be noted only briefly here. Index figures were suggested by Nockolds (1932) as signs of hybridization, but, surely, fragments of quartzite in gabbro might make a hybrid that reduced the index, whereas basalt fragments in granite form a hybrid with a higher index. True hybridization in connection with inclusions involves an admixture of magma or some of its constituents with the solid or plastic inclusion, according to Adams and Barlow (1910, p. 78), Hall and Gardthausen (1911), Goldschmidt (1921), Bailey (1924), Tyrrell (1928), Brammall and Harwood (1932), and Nockolds (1935-1937); or a contamination of a limited volume of magma, either by reaction with the inclusion—as discussed by Winchell (1914), Wilson (1918), Clapp (1921, p. 115), Thomas (1922), Wells and Wooldridge (1931), Thomas and Smith (1932), Thomson (1935), and Kennedy and Read (1936)—or by strewing of crystals from the in-

clusion [7] (Plate 1, figure 2). The reaction of magma with a fragment of an early facies as discussed by Broegger (1890), although perhaps not true hybridization, is so similar to the reaction with a xenolith that it may be indistinguishable. And, of course, hybrids formed in a certain igneous cycle may contribute xenoliths to a magma of a later cycle without much chance of being recognized after a second period of reaction. Exchanges of constitutents have been studied at several places. At some, according to Jones (1930), Grout (1933), and Kennedy and Read (1936), little exchange occurred; at others, according to Read (1923), Nockolds (1932, 1933), Brammall and Harwood (1932), and Bailey (1924), exchange was extensive. Probably most inclusions belong in the "reaction inclusion" groups.

Most of the criteria used by geologists who have written of the origin of inclusions are listed as follows:

Sedimentary xenoliths more or less metamorphosed.....List 1
Some hybrids, migmatites, and reaction inclusions.
Igneous inclusions .....List 2
Igneous xenoliths, fragments of older igneous rocks more or less metamorphosed .....List 3
Some hybrids and reaction inclusions.

Cognate inclusions:
- *Fragments of an early facies*.....List 4
- *Fragmented dikes:* either lamprophyres emplaced before the magma was solid, or any dark dike in palingenetic magma....List 5
- *Segregations almost in place*.....List 6

"The residue remaining after the fusion of a considerable mass of pre-existing rock" was mentioned by Teall (1888) but no criteria for such an origin were offered. It seems probable that most such residues are so softened as to be ill-defined schlieren rather than blocky or spheroidal inclusions.

It is noteworthy that there are six lists, not simply two. Most of the discussions in the literature seem to be incomplete in this regard, and give evidence and arguments favoring one origin and opposing a second, without any adequate discussion of the several other possibilities.

Many of the criteria used in the past are vague, incomplete, inaccurate, and mutually conflicting. This is partly because of a lack of realization of the complexity of the problem and partly because of local peculiarities of the outcrops studied. Perhaps a good many of those listed in the table might well be omitted, but they have all been used in some district, and for most of them a reference to the literature is added. Many of the points mentioned have been stated by authors in

[7] On strewing of xenocrysts, *see* Goldschmidt (1912), Bowen (1922, p. 540, 561), Sederholm (1923), Nockolds (1932, p. 448), Richey and Thomas (1932), Tidmarsh (1932), Nockolds (1933, p. 584-7), Brammall and Harwood (1932, p. 207-208), MacGregor (1936), and G. M. Schwartz: *The contrast in the effect of granite and gabbro intrusions on the Ely greenstone,* Jour. Geol., vol. 32 (1924) p. 108.

description of certain kinds of inclusions, without specifically saying that these features are distinctive or useful as criteria of origin. If they are not mentioned as descriptive of the dark spots of other origin, some of them may be included here, although the original author would not today consider them useful criteria. The listing here is made largely in hope of arousing discussion and of clarifying the evidence. It is hoped that no serious misinterpretations have been made in re-wording some of the statements and classifying them under headings. In applying the criteria the original papers should be consulted for a more adequate statement than can be given here. So many of them contradict others that it should be evident that the points suggested need critical and drastic revision. Only a few survive to the close of the present discussion.

LISTS OF CRITERIA SUGGESTED AS INDICATING ORIGIN OF INCLUSIONS

(Numbered notes, under the heading, "Critical Notes on the Criteria Suggested," follow these lists)

LIST I. CRITERIA OF SEDIMENTARY XENOLITHS

(Nearly all are somewhat metamorphosed)

(A). Best criteria

1. The inclusions are mostly near sedimentary wall rocks (Nockolds, 1934). Note 5.

2. Certain peculiarities of minerals or textures may serve to identify a particular source rock. *See* p. 1555.

3. An unknown inclusion is likely to be sedimentary if associated with recognizable sedimentary inclusions (Tilley, 1921). Note 6. This is better if reinforced by the additional evidence of:

   (a) A variety of inclusions occurring in proportions corresponding to the proportions of different sediments in the wall (Nockolds, 1934), (Read, 1935, p. 624-638). Note 5. *See* p. 1560.

   (b) A series of inclusions grading from recognizable sediment to unknown (Ghosh, 1934). Note 7.

4. The inclusions show bedding commonly by bands of different textures (Hall and Gardthausen, 1911) (Grout, 1933). Note 11.

   (a) Especially if the bands have minerals or combinations that are rare in igneous rocks; as, for example, a band mostly made of pyroxene and quartz. *See* page 1555.

   (b) Especially if the bands are too thin to resemble common igneous bands (Grout, 1933).

5. The inclusion has high Al_2O_3; high silica; high ratio of Al_2O_3 to the sum of alkalies and lime; potash greater than soda; or magnesia greater than lime. Combinations are especially significant (Hall and Gardthausen, 1911) (Ghosh, 1927) (Brammall and Harwood, 1932). Certain minor chemical constituents may be useful in identification of fragments of a particular country rock (Brammall, 1933).

6. A series of inclusions may show all stages of assimilation, from recognizable

sediments to schist or hornfels, and to migmatite (Winchell, 1914) (Ghosh, 1934). Note 7.

7. Minerals: strong indicators are cordierite, kyanite, sillimanite, andalusite, dumortierite (Fitch, 1932), graphite, staurolite (Fitch, 1932) (Williamson, 1935); Note 13; less definite are garnets other than spessartite, orthorhombic pyroxenes, corundum, spinel, and some minerals high in lime (Winchell, 1914) (Tilley, 1921) (Bowen, 1922).

8. Biotite schist inclusions are more likely to be of sedimentary origin, produced from clay, than hornblende schist, which is commonly altered basalt (Grout, 1933, p. 1027). Note 14.

9. Helecitic texture is probably more common in inclusions of sedimentary than in those of igneous origin. Note 1.

10. Hornfels texture is common in altered shales (Grout, 1933, MacGregor, 1936) but may form in rocks of igneous origin. In inclusions, it suggests a shale origin in perhaps six out of ten occurrences. Note 3.

(B). Suggested, but not reliable criteria

11. Differential attack by the intrusive on different sediments; shown by field mapping. Compare List II, 14.

12. Sedimentary inclusions have no size limit (Thomas and Smith, 1932). Note 8.

13. Angularity, or slabby form (Thomas and Smith, 1932). Note 9.

14. Sharp contacts (Thomas and Smith, 1932). Note 10.

15. Sedimentary inclusions with their length oriented by flow of the magma host, may have bands in directions independent of the form (Cloos, 1933).

16. A schistosity in the inclusion not in harmony with the flow structure of the host. Note 12.

17. A biotite-rich zone at the outer edge of the inclusion, with plates oriented tangent to the inclusion (Ghosh, 1934).

18. Porphyroblastic and poikiloblastic textures in inclusions suggest sedimentary origin, but can hardly be distinguished from porphyritic and poikilitic textures.

19. Absence of zircon in the inclusion.

20. Minerals differing from those in the host (Lacroix, 1890) (Rogers, 1911).

21. The magma is more likely to be notably contaminated around sedimentary inclusions than around igneous inclusions (Pabst, 1928) (Jones, 1930) (Tidmarsh, 1932). Note 15.

22. Feldspars and other minerals in sedimentary grains in an inclusion may be zoned by magmatic corrosion, but not likely by zonal growth; the distinction is difficult. Note 1.

23. Mafic minerals are said to be coarser than feldspars in sedimentary inclusions (Nockolds, 1934).

Some hybrid inclusions, reaction inclusions, and other mixtures involving sediments. By reaction with the magma, and by addition of magma either in *lit-par-lit* injections or disseminated, and by addition of emanations, all signs of sedimentary origin may ultimately be lost. Perhaps the most persistent criteria are 4, 6, and 8. Analyses of a series may show additions and exchanges.

LIST II. CRITERIA OF IGNEOUS ORIGIN OF INCLUSIONS

(Nearly all are somewhat metamorphosed)

(A). Best criteria

1. Igneous origin is indicated when most inclusions are near an igneous wall rock of similar texture, minerals, and peculiarities of minerals, especially where fragments can be seen in all stages of separation from the walls (Thomas and Smith, 1932). Note 5.

2. A series grading from known igneous inclusions to those of unknown origin (Cross and Purington, 1899) (Grantham, 1928). Notes 7 and 16.

3. Inclusions of anorthosite, peridotite, and anorthite rocks are of igneous origin; nephelite and other feldspathoid rocks are suggestive but less conclusive (Allan, 1914). Note 13.

4. Hornblende schist is more commonly of igneous than of sedimentary origin (Grout, 1933, p. 1027). Note 14.

5. Ellipsoids, amygdules (MacGregor, 1931), glomeroporphyry, and other recognizable igneous structures.

6. Porphyritic texture indicates igneous origin, if it can be distinguished from porphyroblastic texture, but this is not always possible (Clapp, 1921) (Thomas and Smith, 1932, p. 280) (Goodspeed, 1929, p. 161). Note 1. Possibly all large feldspars in fine groundmass may be regarded as suggesting igneous origin, because probably more than half are igneous.

(B). Suggested, but not reliable criteria

7. Even distribution and uniform size of the inclusions (Lahee, 1931) (Nockolds, 1932). Notes 5 and 8.

8. "True igneous consolidation texture" (Eskola, 1914). Note 16. These are not easily distinguished from metamorphic textures. See items 5 and 6. Phenocrysts are probably cited more often than other textural features (Tyrrell, 1916) (MacGregor, 1931; Macgregor, 1936), but metacrysts even of feldspar and quartz may be added to sediments (Clapp, 1921) (Thomas and Smith, 1932, p. 280) (Goodspeed, 1929, p. 161). Note 1.

9. Uniformity of texture (Thomas and Smith, 1932).

10. Inclusions slabby, like fragments of dikes (Clapp, 1921, p. 114-115). See List V.

11. Rounded to oval or ellipsoidal forms in the inclusions (Phillips, 1880) (Lahee, 1931) (Bowles, 1934). Note 9.

12. Rarely more than one foot thick (Thomas and Smith, 1932). Note 8.

13. Grain size, fine, around .25 mm. (Harker, 1909) (Thomas and Smith, 1932); about 1/5 to 1/10 the size of grain of the enclosing rock (Phillips, 1880, p. 21-22) (Pabst, 1928, p. 339). This is clearly not a correct item; see List I, 10. Note 3.

14. Differential attack by the intrusive on different igneous wall rocks, shown by field mapping, may suggest that the one most attacked furnished inclusions; but see List I, 11.

15. The inclusion has a concentration of minerals formed early in the surrounding intrusive. Note 17. The early minerals may be altered, and perhaps strewed in the magma near by, but not dissolved (Iwao, 1936), whereas an inclusion containing minerals later in the reaction series (Bowen, 1922) than the host rock, should be dissolved. Note 2.

16. Feldspars zoned by growth rather than by corrosion. Note 1. The central zone clouded with opaque inclusions (MacGregor, 1931). Note 1.

17. The ratio of biotite to hornblende is said to remain constant even if quartz and feldspar are added to the inclusion (Nockolds, 1935-1937). This is not probable, because hornblende commonly reacts with late magma to form biotite (Bowen, 1922).

18. The igneous rock around an igneous inclusion is less likely to show contamination than that around a sedimentary inclusion (Thomas and Smith, 1932, p. 277). Note 15.

19. The inclusion has a chemical composition like those of igneous rocks (Rogers, 1911) (MacGregor, 1931). This is not reliable (Adams and Barlow, 1910) (Hall and Gardthausen, 1911) (Jones, 1930) (Grout, 1933). Note 2.

20. The inclusion "lacks the characters" of known sedimentary inclusions (Thomas and Smith, 1932), but the characters are hard to state in a form to be useful.

LIST III. CRITERIA OF XENOLITHS OF OLDER IGNEOUS ROCKS

(A). Best criteria

1. Most of the inclusions occur near an igneous wall rock (Adams and Barlow, 1910, p. 121) (Eskola, 1914) (Dixey, 1925) (Pirsson, 1926) (Cloos, 1931) (von Gottfried, 1932). A wide distribution far from the wall is no disproof of xenoliths (Gardiner and Reynolds, 1932). Note 5.

2. Petrographic correspondence to igneous rocks in the nearby outcrops; it may be in structure, texture, rare minerals (Hurlbut, 1935), peculiar character of certain minerals, or in minor chemical constituents (Brammall and Harwood, 1932). Notes 5 and 13.

3. A series of inclusions grading from one of unknown origin to some recognized as igneous (Adams and Barlow, 1910) (de Lapparent, 1913) (Read, 1924), especially if the unknown differs from the known, as would be expected by reaction with the

magma (Bowen, 1922); but these reactions are not always predictable (Nockolds, 1931, p. 320). Note 2. The gradation may be checked by microscopic and chemical studies. Note 7.

4. When the proportions of different kinds of inclusions correspond to the proportions of different igneous rocks in the wall (with similar solubility and specific gravity) the wall rock is the probable source of inclusions. Note 5.

5. Ellipsoids, amygdules, and similar surface igneous structures in the inclusions (MacGregor, 1931). Notes 3 and 16.

6. The inclusions are not associated with, or similar to, the several differentiation facies of the host rock (Pabst, 1928, p. 367).

(B). Suggested, but not reliable criteria

7. One kind of inclusion near one igneous wall rock and decreasing abundance of inclusions with distance (Knopf and Thelen, 1904-1906, p. 237).

8. Expect xenoliths at shallow horizons (segregations at depth) (Grout, 1926, p. 32) (Pirsson, 1926, p. 168).

9. Irregularity in size and shape. Notes 8 and 9. No limit in size (Lahee, 1931); size and angularity greatest near source (Powers, 1915); sharp contacts and more abrupt transitions than segregations.

10. Inclusions zoned by action of the magma (Bailey, 1924, p. 275) (Read, 1924) (Jones, 1930); but sediments have similar zones (Nolan, 1935). Note 15. Zones may be thin, or may be up to 1000 feet thick (Lahee, 1931); if they are thick, they are visible only in the large inclusions, because the reaction in the small inclusions extends all across the inclusion and is no longer a zone. Zones around inclusions may be chilled, or may show signs of melting (Lacroix, 1890), even some glass (Knopf and Thelen, 1904-1906, p. 239) (Richey and Thomas, 1932).

11. Schist inclusions with igneous composition (Harker, 1909). Note 12.

(a) Especially if the host rock is not foliated (Dixey, 1925).

(b) Especially if the internal structure of the schist inclusion transgresses the foliation of the host, even if the length of the fragment is parallel to it (Cloos, 1933).

12. Sugary hornfels texture (Richey and Thomas, 1932) (Grout, 1933) (Joplin, 1935). This is even more characteristic of a shale origin; see List I, 10. Notes 3 and 16.

13. Poikilitic or sieve texture, especially in mafic minerals (Joplin, 1935); porphyritic, if not porphyroblastic; see List II, 6.

14. Feldspar phenocrysts in igneous xenoliths are not zoned (Pabst, 1928, p. 356). Note 1.

15. An inclusion may be an igneous xenolith if it has an essential or abundant accessory mineral that is not in the host, or not in such coarse grains in the host, or if it has a mineral with peculiarities of color, structure, or other quality, different from those of the same mineral in the host.

16. An inclusion may be an igneous xenolith if it lacks an essential mineral of the host.

17. Orthorhombic pyroxene in the inclusion (Bailey, 1924, p. 355) (Richey and Thomas, 1932). See List I, 7. Note 13.

Hybrid inclusions and mixtures involving igneous xenoliths.

By reaction with the magma and addition of magma or its emanations, the minerals, textures, and composition may be much changed. A series showing intermediate steps in the changes is the best sign of origin, item 3. The strewing of mineral fragments from igneous inclusions into a definite zone of the host makes a sort of hybrid. Compare List I, 21, and List II, 18; see also page 1530. A high content of introduced apatite, titanite, and magnetite may be suggestive (Thomas and Smith, 1932, p. 288).

LIST IV. CRITERIA OF COGNATE FRAGMENTS

(Early facies included in a later facies of the same intrusive in the same chamber.)

(A). Best criteria

1. Inclusions with petrographic peculiarities that resemble those of an early facies in the same magma chamber (Grout, 1926) (Sanders, 1929) (Tansley, Schafer, and Hart, 1933). Note 5. If there is no such early facies exposed, but there are minor

satellitic intrusions of similar peculiarity, they also may suggest that an early border facies once occurred in the main chamber (Mayo, 1935).

2. Inclusions are most abundant near a border facies that they resemble, not different in abundance near different wall rocks (Taber, 1913) (Billingsley, 1915-1916). Note 5.

3. Cognate inclusions show much the same minerals (Pabst, 1928, p. 343) (Heyl, 1936) and the same late magmatic reactions (Bowen, 1922) as the minerals of the host, because the two are related and the fragments were still hot when included. A lack of such reaction in the inclusion is a strong sign that the inclusion is not cognate (Bowen, 1922). Note 2.

4. Anorthosite fragments in diabase and its facies, especially where associated with, or grading toward, glomerophenocrysts of plagioclase, are believed to be cognate. Note 13.

5. Olivine in granular lumps in basalt (Phillips, 1880) (Harker, 1909) (Bailey, 1924, p. 369) seems to be cognate but may have no analogue in plutonic rocks. Note 18.

(B). Suggested, but not reliable criteria

6. Inclusions that differ from all recognized xenoliths (Grout, 1926) (Pabst, 1928, p. 356) and are not such as would be expected from known wall rocks by reaction with the magma (Grimsley, 1894-1895) (Knopf and Thelen, 1904-1906).

7. Cognate fragments may be of any size, but few are more than 20 feet thick (Clapp, 1921) (Tansley, Schafer, and Hart, 1933). Note 8.

8. Cognate fragments are more common in intermediate rock than in granite or gabbro (Pabst, 1928, p. 358) (Mayo, 1935). Note 7.

9. Cognate fragments are not associated with known segregations of similar character (Clapp, 1921, p. 46). Note 6. See List VI.

10. Cognate fragments are not part of a series grading to known metamorphosed sediments (Pabst, 1928, p. 358) (Mayo, 1935). Note 7.

11. A dark cognate inclusion might darken the host rock near it, in contrast to a dark segregation which might leave a zone free from dark minerals (Pabst, 1928, p. 341) (Nockolds, 1933) (Heyl, 1936). Note 15.

(a) Many are rounded as if by attack, some angular (Phillips, 1880) (Billingsley, 1915-1916) (Clapp, 1921) (Williams, 1932). Note 9.

12. Foliation in a cognate inclusion is a result of igneous flow, not a schistosity with luster from cleavage of mica (Pabst, 1928, p. 356) (Williams, 1932, p. 205). See page 1566 of this paper, on Giants Range. Note 12.

13. "True igneous textures," granitoid, porphyritic. Note 16. Compare with List III, 13.

(a) Reaction may give poikilitic and replacement textures.

(b) Some fragments of early border facies may be finer grained than late facies, with grain 1/5 to 1/10 that of the matrix (Harker, 1909). See page 1564 of this paper, on Alta stock. See List I, 10.

14. The difference between intrusive and a cognate inclusion suggests the normal differentiation trend rather than a trend like that to lamprophyres. See List VI, 11.

15. A cognate inclusion has a concentration of the early minerals of the host. Note 17.

16. A cognate inclusion has the same mineral peculiarities—inclusions, color, haloes,—as in the host (Bowles, 1934).

17. Cognate inclusions have no orthorhombic pyroxene and no pyrogenic hornblende (Bailey, 1924, p. 355). Note 13. Compare List III, 17.

18. Hybridism indicates cognate origin (Harker, 1909, p. 336). This is denied (Clapp, 1921, p. 126).

LIST V. CRITERIA OF FRAGMENTED DIKES

(Either lamprophyres that have been fragmented before the surrounding intrusive cooled (Note 4), or any dikes in palingenetic magma or magma replacing easily replaced rock)

(A). Best criteria

1. Gradation from fragments to dikes (Adams and Barlow, 1910), p. 121) (Clapp, 1921) (Dixey, 1925).

2. Abundant basic fragments in aplite near the border of an intrusive (Sederholm, 1923, p. 130, and pl. 6) (Balk and Grout, 1934). *See* Plate 6, figure 2 of this paper.

3. Lamprophyric rock types in fragmented slabby inclusions (Daly, 1912-1913, p. 349) (Sederholm, 1926, p. 18) (Balk and Grout, 1934).

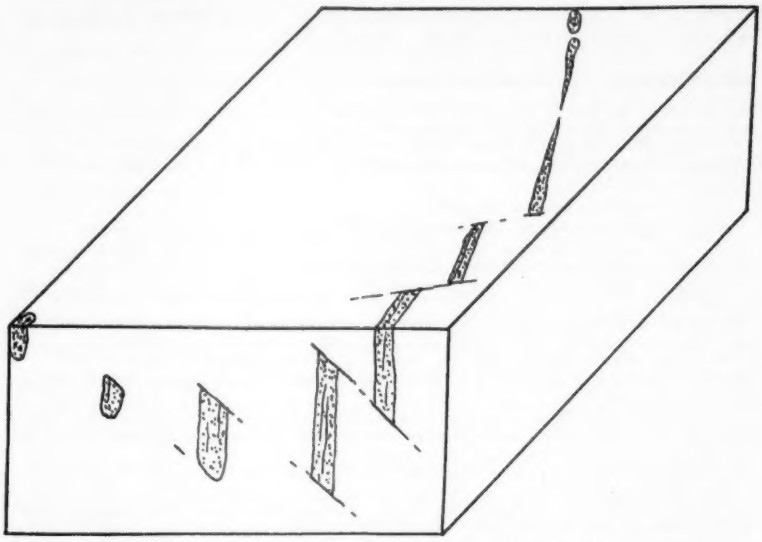

FIGURE 1.—*Generalized sketch of field evidence that some inclusions are fragmented dikes*

Where the fault planes are healed by residual magma, or by thorough recrystallization, the evidence of origin may be wholly lost.

(B). Suggested, but not reliable criteria

4. Old-looking gneiss as the host for inclusions (Sederholm, 1926).

(a) Various evidences of deep, high-temperature alteration in the host.

(b) Gradation of the intrusive host into a metamorphic rock such as conglomerate (Balk and Grout, 1934, pl. 87, A).

5. Slabby inclusions differing from slabs that could be derived from wall rocks (Smyth and Buddington, 1926).

6. Slabs in a replacement type of intrusive, where the slabs are oriented in harmony with the dikes in the wall rock (Callaghan, 1935).

LIST VI. CRITERIA OF SEGREGATION

(Almost in place, from the surrounding intrusive, into forms like inclusions)

(A). Best criteria

1. Texture should be granitoid or coarse, because segregation is not likely to occur where cooling is so rapid that the rock is fine grained. Note 16.

2. Glomeroporphyritic clusters have probably segregated (Harker, 1909) (Bailey, 1924, p. 290) (Tyrrell, 1928, p. 132). Note 16.

3. Orbicules may have segregated, especially if they do not have xenoliths at the centers (Gilbert, 1906) (Adams and Barlow, 1910, p. 139) (Knopf, 1918) (Lahee, 1931). Zones formed by reaction of xenolith and magma may be distinguished with difficulty (Bowen, 1922) (Pabst, 1928, p. 341). Note 15.

4. Segregated lumps, containing early minerals of the host, show the same late magmatic reactions (Bowen, 1922) as the minerals of the host, because the lumps are as hot as the rest when reaction occurs. Notes 2 and 13.

(B). Suggested, but not reliable criteria

5. An igneous inclusion, in a host that has no exposed border facies or wall rock of the sort, may be segregated (Pabst, 1928, p. 367) (Reid, 1912). Note 5.

6. The segregation differs from known xenoliths (Pabst, 1928, p. 356). Compare List IV, 6.

7. Segregations are uniform in character and distribution (Cloos, 1931, p. 379) (Miller, 1934, p. 31). Note 5. Compare List II, 7.

(a) Not different near different walls.

(b) Not all near the roof (Emmons, 1926, p. 427); or walls (von Gottfried, 1932).

8. If a rock has basic segregated inclusions, its basic facies may have similar more basic segregations, but not xenoliths (Pabst, 1928, p. 340, 367). Note 7.

9. Segregated inclusions do not form a series grading to other known rock xenoliths (Pabst, 1928, p. 358). Note 7.

10. The intrusive shows a differentiation series, at borders or in schlieren, to rocks like the inclusion (Adams and Barlow, 1910, p. 121) (Daly, 1912-1913, p. 349) (Wagner, 1914) (Bowles, 1918, p. 85) (Clapp, 1921) (Billingsley, 1915-1916). Note 7. This is the opposite of item 5.

11. A segregation may show a differentiation trend like that to lamprophyres (Pabst, 1928, p. 362). Compare List V, 3.

12. Gradation from inclusion to the host is common in segregations (Daly, 1903b, p. 43). Compare List I, 21. Note 7. The gradation may form masses that are concentrations, up to 100 per cent, of each mineral of the host (Cross and Purington, 1899).

13. Segregations are more common in intermediate rock than in granite or gabbro (Pabst, 1928, p. 337). Compare List IV, 8.

14. Forms rounded, oval, or slabby, not angular (Phillips, 1880) (Lahee, 1931) (Stansfield, 1928, p. 27) (Emmons, 1926) (Bowles, 1934). Note 9.

(a) Lack of angularity is not a good criterion, but the presence of angularity is good evidence that the lump is not segregated in place (Phillips, 1880) (Williams, 1932).

(b) Slabby segregations are parallel to the structure of host (Gilbert, 1906) (Cloos, 1933) (Knopf and Thelen, 1904-1906).

15. Segregated lumps have a structure parallel to that of the host (Cloos, 1933). Note 12. Internal structure is a result of igneous, not metamorphic flow; based more on feldspar orientation than on biotite schistosity and luster (Daly, 1903b). A lack of flow structure in the host suggests growth of segregation after emplacement (Daly, 1903b). Note 12.

16. Segregated lumps are not large; few reach a foot in diameter, but they are commonly uniform in size in a particular outcrop (Lahee, 1931). Note 8.

17. Segregations may have a texture that is sugary, or porphyritic with sugary groundmass (Phillips, 1880) (Bastin, 1908) (Harker, 1909) (von Gottfried, 1932) (Knopf, 1918). Note 3. They are finer than the host because early minerals to crystallize form small crystals (Harker, 1909).

18. Crystals may cross the contact from segregation into host (von Gottfried, 1932) (Knopf, 1918) (Pabst, 1928, fig. 3); or enclose sugary grains poikilitically.

19. If an inclusion is not such material as the walls or early facies could furnish by inclusion and reaction, it is more likely to be a segregation (Grimsley, 1894-1895) (Knopf and Thelen, 1904-1906) (Bowen, 1922). Note 5. Reaction, however, may cause surprising changes.

20. Segregations are not so dark as xenoliths (Pabst, 1928, p. 356).

21. Minerals in segregations are the same as in the host (Phillips, 1880) (Grimsley, 1894-1895) (Bowles, 1918) (Lahee, 1931) (Harker, 1909, p. 322) (Bowles, 1934). Note 13. Even the accessories are the same (Knopf and Thelen, 1904-1906) (Pabst, 1928, p. 358). The segregation may be rich in the early minerals of the host—sphene, apatite, magnetite (Knopf and Thelen, 1904-1906) (Knopf, 1918). This is safe only when early accessories can be distinguished from late additions (Thomas and Smith, 1932, p. 288). Note 17.

22. Segregated minerals have the same peculiarities of color, haloes, and inclusions as those in the host (Bowles, 1934).

23. Feldspar is less abundant in segregated lumps than in the host but may be like that in the host; or the segregations may lack the orthoclase and quartz of the host (Phillips, 1880) (Knopf, 1918) (Pabst, 1928, p. 358). Note 1.

CRITICAL NOTES ON THE CRITERIA SUGGESTED

GENERAL STATEMENT

Several of the assumptions in the criteria listed have been subjects of so much disagreement as to need a few pages each in critical summary; after which a number of minor criteria are briefly noted as to their probable usefulness or lack of it. For some criteria the discussions in the papers referred to in the lists of criteria are sufficient, and space is not needed here for criticism.

NOTE 1. FELDSPATHIZATION AND FELDSPAR "PHENOCRYSTS" INCLUSIONS

Harker [8] has recently summarized the relation of feldspars to metamorphism, both thermal and pneumatolytic. In several chapters he makes it clear that albite grains, even porphyroblastic albite, may result from recrystallization without essential additions. In writing of feldspathization, he refers to "the supposed impregnation on an extensive scale of metamorphosed rocks with felspar from a magmatic source. This hypothesis dates from a time when the formation of various felspars as normal products of metamorphism was not adequately appreciated. The proof of addition of material to a rock should naturally be sought in a series of comparative chemical analyses, but such evidence is seldom forthcoming." [9] He notes the Stavanger district (Goldschmidt, 1921) as an example in which soda and silica (not albite) were added but did not penetrate far from the magma. In a general study of metamorphism, this is no doubt the right attitude to take, but in a study restricted to inclusions, mostly of small size, in magmas of large size, the point of view is notably different. Magmatic emanations would not have to travel far to permeate an inclusion thoroughly, and the chemical evidence is available to prove that in a number of districts they have done so. (*See* the next heading.) Furthermore, the field evidence that feldspar metacrysts are added, may at some places be convincing, even when the analyses of bedded material like shale leave the conclusion uncertain. Plate 15 shows granite working its way into a sedimentary inclusion.

This is a matter of considerable importance, and in view of a radical difference of opinion expressed in earlier papers several specific examples are here cited. Tyrrell (1916) says phenocrysts are a sign of igneous origin even when associated with minerals common in altered sediments,

[8] Alfred Harker: Metamorphism (1932) p. 55-56, 212.
[9] *Op. cit.*, p. 250.

such as scapolite and epidote. Thomas and Smith (1932) "could not accept the growth of phenocrysts in xenoliths" at Tregastel, and believe that rocks at Shap, at Flamanville, and at Peterhead are so like those at Tregastel as to suggest igneous origin. The earlier work at Shap also attributed the feldspar phenocrysts to early formation, but Grantham (1928) found evidence of late additions of TiO_2 and the appearance of added feldspar. MacGregor (1936) found hornfels from shale in the Dalbeattie pegmatite, but where similar rocks have phenocrysts he refers them to a porphyry in the country rock.

On the other side of the argument there are not only Stavanger and the Giants Range, but a number of others. The slates in the Duluth gabbro have large additions of feldspar or the constituents to produce feldspar; chemical as well as mineralogical studies are available (Grout, 1933). The rocks of Essex County, Massachusetts, show evidence of added phenocrysts (Clapp, 1921). Inclusions in the granite of the Rumney quadrange, New Hampshire, have added feldspar.[10] Analyses as well as field observations show that the quartzite masses of the Bushveld norite are feldspathized (Daly, 1928). The rocks of Cornucopia, Oregon, have plagioclase "phenocrysts" grown in xenoliths (Goodspeed, 1929). Schist xenoliths in the Inverbeg lamprophyre are said to be feldspathized.[11] The inclusions in granite at Bibette Head received late phenocrysts or porphyroblastic crystals (Nockolds, 1932). Feldspar metacrysts grew during contact action by the batholith at Cassia, Idaho.[12] The large inclusions in the igneous rocks of Jersey have developed outer porphyritic zones by reaction with the magma (Wells and Woodridge, 1931). A large granite mass in the Inyo Range has probably resulted from granitization and albitization.[13] The quartzites north of Lake Huron have been changed to feldspathic gneisses (Collins, 1935). Large porphyroblasts of feldspar were added to xenoliths in the Kuga district, and the additions are established by the fact that the outer zones are more sodic than the outer zones of the plagioclase of the granite host (Iwao, 1936). Graywacke inclusions in the dike of Newmains are feldspathized (Kennedy and Read, 1936).

Albitization as a late magmatic process should be fully as effective on small inclusions as on the main or border phases of an intrusive. The

[10] Lincoln Page: *The geology of the Rumney quadrangle, New Hampshire*, Univ. Minn., thesis (1937).

[11] Sir John Flett: *The Inverbeg lamprophyre*, Geol. Surv. Great Britain, Summ. Prog. 1928, pt. 2 (1929) p. 29-35.

[12] A. L. Anderson: *Contact phenomena associated with the Cassia batholith*, Jour. Geol., vol. 42 (1934) p. 376-392.

[13] G. H. Anderson: *Granitization, albitization and related phenomena in the northern Inyo Range of California-Nevada*, Geol. Soc. Am., Bull., vol. 48 (1937) p. 17-74.

literature of albitization and related processes has been summarized with references by Gilluly [14] and need not be discussed here except to say that a feldspar composition more than 90 per cent albite, especially if more than 96 per cent albite, is strongly suggestive of introduction; and certain replacement textures such as chessboard twinning and even graphic patterns are also suggestive of introduction.

The extreme of these ideas is the suggestion that not only the coarse feldspars of the schist but the phenocrysts of the granite itself (Pl. 1, fig. 3) result from replacement. There are in the literature perhaps a dozen descriptions of rocks with porphyritic borders where late phenocrysts may have grown.[15]

Taking all the evidence, together with some given under the heading on chemical data, there seems to be little reason to doubt that feldspars may be added to many xenoliths, that some (whether called phenocrysts, metacrysts, or porphyroblasts) may have coarser grain than the bulk of the xenolith, and that some replace the material of the xenolith.

With this in mind the suggested criteria can be judged somewhat critically. Certainly the occurrence of coarse feldspar is no proof at all of an igneous origin. If a medium or basic feldspar occurs in an inclusion, it may be zoned with higher percentages of albite outside, regardless of the origin of the inclusion; zoning is of little value as a criterion (Goldschmidt, 1912) (Bowen, 1922) (Iwao, 1936). If the zoning that results from growth can be distinguished from zoning by later corrosion, it is a useful sign. The "clouding" of feldspar is believed by Macgregor (1931, p. 524-531) to be a sign of thermal metamorphism of a feldspar already solidified, but clouded feldspars have formed where feldspars grew during thermal metamorphism of a slate (Grout, 1933, p. 1014, 1020, 1032) (Pl. 2, fig. 2). It cannot safely be used as a criterion of feldspathic original rock.

This rather sweeping rejection of the features of feldspars as proofs of igneous origin should perhaps be qualified. Many useful criteria are not strict proofs. Probably there are more igneous rocks than altered sediments that have coarse feldspars. Together with other criteria, they may suggest igneous origins for inclusions, but it is believed that the suggestion is not strong—perhaps six times out of ten rather than nine times out of ten.

It seems likely that a helecitic structure would appear in coarse feldspars formed by replacement of schist xenoliths. This would be satisfactory evidence of replacement; but up to this time it does not seem

[14] James Gilluly: *Replacement origin of the albite granite near Sparta, Oregon*, U. S. Geol. Surv., Prof. Paper 175c (1933) p. 69-80.

[15] *See*, for example, A. L. Anderson: *op. cit.*

to have been recorded as a criterion of origin of feldspar in xenoliths. (*See* description of Giants Range, to follow.) (Pl. 13, fig. 2.)

It should be noted that several geologists have assumed extensive replacement in the walls of an intrusive without giving a satisfactory list of criteria or discussion of the evidence.[16] To most petrographers the term replacement, used in description of a rock, brings to mind a process of metasomatism. The use of "replacement" for large-scale processes of crowding aside a schist, with some assimilation and lit-par-lit injection, is confusing.

Finally, it may be noted that quartz may be added to inclusions in a few districts in much the same way as feldspar (Daly, 1912-1913, p. 526-528) (Pabst, 1928, p. 342) (Grout, 1929, p. 574, 583) and may form metacrysts.[17]

NOTE 2. CHEMICAL DATA AS CRITERIA

Analyses of inclusions are often inspected to find evidences of origin, but the results are dependable only in one direction. If the signs of sedimentary origin are conspicuous the evidence is conclusive; Al_2O_3 greater than the alkalies and lime, magnesia greater than lime, potash greater than soda, and other less definite features.[18] If these signs are not found, it is not safe to conclude anything as to the origin of inclusions, for reaction with a magma commonly changes them so that chemical as well as mineralogical signs of sedimentary origin are lost. There is nothing in the igneous character of an analysis of an inclusion to warrant the assumption of igneous origin. Probably half the inclusions in granites and granodiorites are sediments so far altered as to be identified with difficulty.

Analyses of material of known sedimentary origin but with compositions well within the range of common igneous rocks have been reported from several districts. The slate hornfels inclusions in the Duluth gabbro are of gabbro composition (p. 1556). Those of the Bushveld are analogous (Hall and Gardthausen, 1911). Even as early as 1888, Williams had shown by analyses that schist of sedimentary origin grades, as a result of contact action, into rocks of a composition which looks igneous. Very different from these examples of gabbro action on clays is the action of granite on limestone, but the amphibolites so formed

[16] For examples *see* J. T. Stark: *Migmatites of the Sawatch Range, Colorado*, Jour. Geol., vol. 43 (1935) p. 1-26; T. L. Kesler: *Granitic injection processes in the Columbian quadrangle, South Carolina*, Jour. Geol., vol. 44 (1936) p. 32-44; and D. L. Reynolds: *Demonstrations in petrogenesis from Kiloran Bay*, Mineralog. Mag., vol. 24 (1936) p. 382.

[17] G. E. Goodspeed: *Development of quartz porphyroblasts in a siliceous hornfels*, Am. Miner., vol. 22 (1937) p. 133-138.

[18] C. K. Leith and W. J. Mead: *Metamorphic geology* (1915) p. 226-240.

E. S. Bastin: *Chemical composition as a criterion in identifying metamorphosed sediments*, Jour. Geol., vol. 21 (1913) p. 193-201.

For examples of use on inclusions *see* E. B. Bailey (1924, p. 34) and Ghosh (1927).

have compositions like those of igneous rocks (Adams and Barlow, 1910, p. 108-127).

Harker, as already mentioned, reports that chemical analyses as proof of metamorphic additions are seldom given, but he cites, in addition to the Stavanger case, eight or ten examples of pneumatolytic additions of soda. Several other papers cite evidence of additions to inclusions. Analyses of clay and hornfels show large additions of soda in one rock of the Kristiania district (Goldschmidt, 1912, p. 36-37). In Mull, there is clear evidence of the modification of xenoliths by magmatic matter (Thomas, 1922). At Bodmin Moor, inclusions are shown by analyses to be made nearly like igneous rocks, but they still have some sedimentary characters (Ghosh, 1927). The Shap granite inclusions have TiO_2 added to their margins and may be modified more generally (Grantham, 1928). The inclusions of quartzite at Sudbury, Ontario, show changes indicating magmatic additions (Jones, 1930). In the Alpen-Ostrand the chemical analyses indicate changes of composition of the inclusions (Schoklitsch, 1935). The slate inclusions in the Carnmenellis granite are progressively enriched in alkalies (Ghosh, 1934). The pegmatite of the Dalbeattie granite penetrates the inclusions, adding soda and probably other elements (Macgregor, 1936). At Barnavave, there seems to have been addition of quartz and silicic feldspar to inclusions (Nockolds, 1935-1937). Read has emphasized the generality of an exchange of material between inclusion and magma (Read, 1923), and several other petrographers find similar evidence (Read, 1935), but Kennedy and Read (1936) found that some potash was added to a hornfels without exchange in the dike of Newmains. Finally, in several districts where the writer has met the problem and suspected from field relations that the composition had changed, the rocks have been analysed and additions were found, some of them probably as an exchange reaction, but others in contact rocks where there were no contributions from the slate to the magma.

Taking all these factors into account, there can be little doubt that a magma may so change an inclusion, while it still remains an inclusion, that the chemical criteria of origin are lost. This is a perfectly logical conclusion from Bowen's careful study of the physical chemistry of the inclusions in magmas (1922), as well as from the descriptions of many districts. All inclusions in granitoid igneous rocks should be expected to have igneous compositions except those included at so late a magmatic stage that they had no time to react. It is possible, of course, that some inclusions might react to equilibrium, but have a proportion of minerals that would be rare in igneous rocks; as, for example, a greenstone, turned to a biotite-rich inclusion in a biotite granite.

Notwithstanding this increasing realization of magmatic reaction on inclusions, there have been recent papers in which the inclusions are described as igneous, without any adequate discussion of the reason for such an assumption. At Tregastel the field evidence may be sufficient without the chemical evidence of igneous origin, but when Thomas and Smith (1932) extend their suggestions to similar inclusions in other districts the field evidences need a careful check before being accepted. MacGregor (1931) had good evidence (amygdules) that Scottish pyroxene granulites were of igneous origin, but when he attributes Odenwald beerbachites to igneous origin because of petrographic similarity and igneous composition, his remarks are not at all conclusive. At Barnavave, Nockolds has xenoliths that he believes analogous to pyroxene granulites, but his paper (1935-1937) gives no facts to show that they could not be altered shales. Pabst (1928, p. 368) makes use of analyses as indicating "autoliths" in the Sierras, but analyses give only the weakest of evidence, and the texture is suggestive of a hornfels.

It should now be well recognized on the basis of 40 years' experience,[19] as well as on the basis of physico-chemical data (Bowen, 1922) that metamorphism by a magma may make a xenolith resemble a cognate fragment so closely that it loses all chemical signs of parentage.

Chemical work found a different application at Dartmoor (Brammall, 1933), the minor constituents (BaO) in inclusions occurring in amounts characteristic of one source rock and not another.

The comparison of analyses of a series of rocks is commonly done very loosely, on the assumption of constants that are not probable. Nockolds (1935-1937) called attention to the danger in this procedure and used more accurate methods, involving specific gravity, porosity, and the assumption that the altered rock occupied the same volume as the original.[20] This one assumption is reasonable, because the rocks are crystalline and not very porous. The possible solution of the outside, and the strewing about of xenocrysts from the inclusion, does not affect the assumption of constant volume for the part left in the inclusion.

For three districts described in later parts of this paper, where chemical data are available, such calculations have been made; Table 1 is a summary of the results. It is believed that all errors of assumptions and data are so slight that the tabulated results are nearly correct—probable errors no more than 5 per cent of the reported gains and losses.

[19] Beginning at least as early as a paper by G. H. Williams on the general relations of the granitic rocks in the middle Atlantic piedmont plateau [U. S. Geol. Surv., 15th Ann. Rept. (1893) p. 662-664].

[20] *See* W. Lindgren: *Contact metamorphism at Bingham, Utah*, Geol. Soc. Am., Bull., vol. 35 (1924) p. 533.

Where data are sufficient, averages are used instead of single analyses. Some minor constituents have been omitted from the calculation.

Direct magmatic additions are indicated where the porosity is reduced and where some constituents have been lost. If the changes took place after inclusion in the magma, an exchange is indicated, but much change

TABLE 1.—*Gains and losses, in grams per 100 cc., by shales and slates included in three intrusives*

(Calculated on basis of equal volumes. Data from Tables 3, 5, and 6)

	Duluth gabbro			Boulder Batholith			Giants Range		
	2 Slate and Graywacke	3 Hornfels inclusions	Change	3+ Shale and Phyllite	3+ Hornfels and "Porphyry"	Change	1 Schist	1 "Porphyry"	Change
Sp. gr.	2.747	2.864	+.117	2.719	2.862	+.143	2.735	2.862	+.127
Porosity	1.57%	1.11%	−.46%	.84%	1.12%	+.28%	.76%	1.00%	+.24%
SiO_2	156.6	140.4	−16.2	182.0	147.5	−34.5	174.5	151.6	−22.9
Al_2O_3	50.8	49.7	−1.1	42.5	54.3	+11.8	43.6	44.3	+.7
Fe_2O_3	2.9	4.9	+2.0	1.4	7.4	+6.0	1.9	5.3	+3.4
FeO	15.9	19.3	+3.4	11.8	17.7	+5.9	12.4	16.3	+3.9
MgO	11.7	21.1	+9.4	6.2	13.0	+6.8	8.8	19.6	+10.8
CaO	6.9	33.8	+26.9	7.6	23.6	+16.0	7.0	20.2	+13.2
Na_2O	7.6	5.8	−1.8	3.6	7.6	+4.0	8.4	11.9	+3.5
K_2O	5.4	3.0	−2.4	6.0	3.8	−2.2	6.9	4.5	−2.4
H_2O+	7.7	1.8	−5.9	3.8	2.3	−1.5	3.9	4.4	+.5
CO_2	1.1	.9	−.2	.....	.....	.....	.....	.....	.....
TiO_2	2.1	2.6	+.5	1.7	2.8	+.9	1.5	2.5	+1.0
P_2O_5	.2	.3	+.1	.7	.9	+.2	.6	1.0	+.4
MnO	.1	.2	+.1	.2	.5	+.3	.2	.4	+.2
Total gains			+42.4			+51.9			+37.6
Total losses			−27.6			−38.2			−25.3
Balance gained per 100cc.			+14.8			+13.7			+12.3

may occur in contact zones, where additions come from the magma but losses are carried away by passing solutions.

Table 2 compares these gains and losses in the inclusions of three districts with similar gains and losses in other districts. Argillaceous rocks in basic magmas commonly lose silica and alumina and gain ferrous iron, lime, and magnesia, becoming in all these respects more like the rocks in which they are enclosed. Argillaceous rocks in more silicic magmas gain soda but are less regular in other changes. It is clear that changes may make the sediments resemble igneous rocks in chemical composition, so that chemical criteria of the igneous origin of any inclusions should be used only with the greatest caution. The

TABLE 2.—*Gains and losses of constituents by inclusions in plutonic igneous rocks of certain districts, as reported by authors*

Districts	Constituents													
	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO−	Na_2O	K_2O	H_2O+	CO_2	TiO_2	P_2O_5	MnO	FeS−
In Gabbro, diorite, and tholeiite														
*Shale, Duluth	−	−	+	+	+	+	−	−	−	−	+	+	+	
Shale, Bushveld (Hall and Gardthausen, 1911)		−			+	+								
Clay, Mull (Thomas, 1922)		−		+	+	+		+	−					
Clay, Mull (Bailey, 1924)	−	0	+	+	+	+		+						
Clay, Arnage (Read, 1923)		−			+	+		−						
Slate, Huntley (Read, 1924)		−			+	+	−	−						
Schist, Peekskill (Williams, 1888)	−	+	+	+			−	−						
In granite, tonalite, and syenite														
*Slate, Giants Range	−	+	+	+	+	+	+	−	+		+	+	+	
*Shale, Boulder	−	+	+	+	+	+	+	−	−		+	+	+	
Clay, Kristiana (Goldschmidt, 1912)							+							
Clay, Stavanger (Goldschmit, 1921)	+						+		−					
Clay, Loch Awe (Nockolds, 1934)	+	−		−	−			+						
Graywacke, Newmains (Kennedy and Read, 1936)							+	+						
Shale, diabase, Dartmoor (Brammall and Harwood, 1932)	+	0	−	−	−	−	+	+			−	+	−	
Greenstone, Dhoon (Nockolds, 1931)		+		−	−	−	+	+						
Dolerite, Barnavave (Nockolds, 1935–1937)	−	0	−	−	−	−	+	+			−	+	−	
Limestone, Haliburton (Adams and Barlow, 1910)	+	+	+	+	+	−	+	+		−	+			
Limestone, Loch Awe (Nockolds, 1934)							+							
Limestone, Strath Oykell (Read, 1926)						−		+						
**Limestone, Bingham	+	+?		+	+	−?	+			−				+
Limestone, Dorback (Walker and Davidson, 1935)		−	+	+	+	−								
Limestone, Barnavave (Nockolds, 1935–1937)	+	+	+	+	−	−				−	+	+	+	

* Data in this paper, in section headed "Inclusions in certain significant masses."
** See reference page 1549.

data also confirm the conclusion of Nockolds (1934) that gains and losses are not uniform, but differ with the magma, with the sediment, and with other factors (Read, 1935, p. 626).[21]

[21] A complex relation of alumina to magmatic action is suggested by A. Brammall [*The reciprocal rôle of alumina in reaction series*, Mineralog. Mag., vol. 24 (1936) p. 362-366].

NOTE 3. SUGARY, HORNFELS TEXTURE IN INCLUSIONS

The hornfels texture, which is commonly formed in a solid fragment or wall rock by contact metamorphism, is characteristic (Pls. 2, 3, 4) and is not common in igneous rocks, being limited almost wholly to "diaschistic" dikes (Pl. 4). This texture has been given almost as many names as the inclusions that show it: hornfels, granulitic mosaic, decussate, muscovado, granoblastic, sugary.

A paper by G. A. Joplin (1935) calls attention to the fact that a great variety of materials when subjected to contact action converge to this textural type. Not only shales and basalts but even granitoid rocks become sugary grained. In its essentials the texture seems to result from simultaneous recrystallization of a group of minerals, without any pronounced difference in their "crystallizing power" and without differential stress.[22] All growing at once, they yield a mosaic of grains of almost equal size. The common size range is from 0.1 to 1 millimeter, and typical hornfels has grains about 0.5 millimeter. If some minerals such as biotite or feldspars grow large, the texture becomes poikilitic with the little granules still about the typical size. On the other hand, if a rock consists almost wholly of one mineral, such as quartz, or iron oxide, the same degree of contact action commonly produces a much coarser product. The determining factors thus seem to be (1) recrystallization and (2) a group of at least two or three minerals. It is noteworthy that the texture becomes sugary almost regardless of the original texture. The same magma that commonly turns a slate to a coarser sugary hornfels makes a granitoid rock finer grained and sugary (Pl. 5, figs. 1, 2). Examples are too numerous to doubt (Ghosh, 1927) (Wells and Wooldridge, 1931) (Grout, 1933) (Joplin, 1935).

Two of Joplin's suggested interpretations, however, find less support in studies of other inclusions. There may be a transition from granitoid to hornfels textures, with an intermediate poikilitic texture (Pl. 5, fig. 2), but poikilitic texture can be developed from recrystallization of an originally fine-grained rock as well as from a coarse one (Pl. 5, figs. 2, 3). And again, the suggestion that sugary xenoliths, abundant in acid and intermediate rocks, result mostly from early basic facies bears little weight, because abundant hornfels inclusions are found in gabbros also. Inclusions in gabbros have not been described from so many districts as inclusions in granites, probably because there are not so many large masses of gabbro as there are of granite. The occurrence of hornfels in gabbro, however, should not be attributed to a lack of hybridism but, rather, to the fact the gabbros and other intrusives with hornfels inclusions have not been so viscously deformed as many granites, which have dragged their inclusions and wall rocks into schists (Grout, 1933, p. 1025-1026).

[22] Alfred Harker [*Metamorphism* (1932) p. 34-5] describes and explains decussate structure.

A review of the literature makes the facts in the matter fairly clear. The hornfels texture indicates nothing except a recrystallization at some time. When statistics are available it may be found that such hornfelses are derived from altered shale in considerably more than half the known occurrences, but this is yet too uncertain to be very valuable as a criterion of sedimentary origin. It is suggestive (List 1, 10).

Any sugary-textured inclusion in an igneous intrusive should, in the absence of aplite and lamprophyre, be considered a recrystallized xenolith and not a segregation. This probably means that about nine out of ten sugary inclusions are xenoliths. On the strength of this conclusion, many of the "segregations" mentioned in published papers are seriously questioned. The sketches and photographs accompanying the descriptions indicate strongly that the material has recrystallized forming hornfels; that it is not a segregation, and probably not a cognate fragment, but almost certainly a xenolith; but the xenolith is of uncertain original nature (Holland, 1900) (Daly, 1912-1913, p. 346-349) (Knopf, 1918) (Dale, 1923) (Martin, 1925) (Pabst, 1928) (Sugi, 1930) (Jeremine, 1931) (von Gottfried, 1932) (Richey and Thomas, 1932) (Nolan, 1935) (Smith, 1936). Half of these have been described as segregations, since the publication of Bowen's discussion. On the other hand, several writers describing British granites (Nockolds, 1932) (Thomas and Smith, 1932) (Brammall and Harwood, 1932) have wisely dropped all reference to segregation, or state specifically that the objects that the older writers called segregations are xenoliths.

A complex problem in rocks of this texture was furnished by the Odenwald granulites, an easily available discussion of which was issued by Macgregor (1931). The rocks were originally described as dikes and were given a distinctive name (beerbachite) on account of their characteristic texture. Klemm found them to be inclusions which he thought were of sedimentary origin. Macgregor compared them with Scottish hornfelses derived from basalt (proved by the fact that amygdules persisted through alteration), and found them so nearly identical that he felt sure the Odenwald rocks must have been derived from fine-grained igneous originals. Tyrrell considered the evidence—textural, mineralogic, and chemical—perfectly conclusive as to the origin from fine basic igneous rocks.[28] It must be agreed that the data for Scottish rocks are perfectly conclusive, but, so far as Macgregor reports the data for Odenwald granulites, there is nothing to determine whether the original was basalt or clay. Both rocks are readily altered to hornfels by gabbro intrusives, and the two products are not easily distinguishable by texture, minerals, or chemical composition. (*See* later notes on Duluth gabbro xenoliths.)

[28] G. W. Tyrrell: *Science progress* (1933) p. 589.

FIGURE 1. GRAYWACKE, X40.

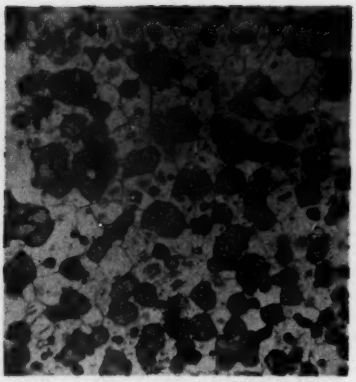

FIGURE 2. COMMON HORNFELS, X40.

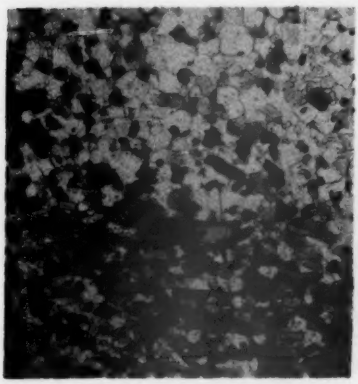

FIGURE 3. HORNFELS RETAINING BEDS OF DIFFERENT COMPOSITION, X50.

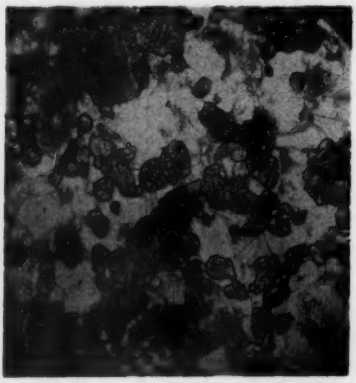

FIGURE 4. HORNFELS FROM CONGLOMERATE, X40.

PHOTOMICROGRAPHS OF GRAYWACKE SLATE AND HORNFELSES FORMED FROM SEDIMENTS BY ACTION OF THE DULUTH GABBRO MAGMA

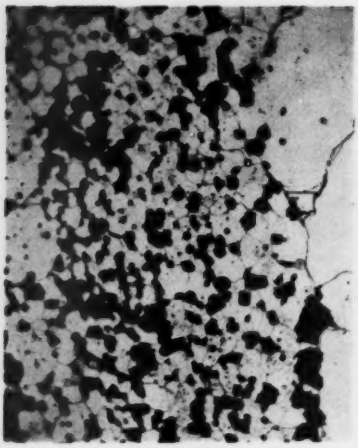

FIGURE 1. FROM IRON FORMATION, X30.

FIGURE 2. FROM PORPHYRY, X20.

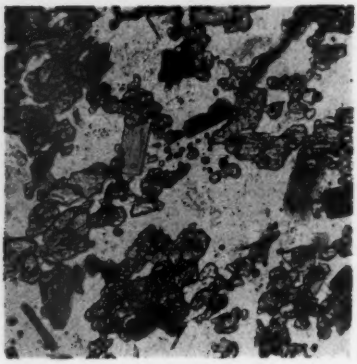

FIGURE 3. FROM MICA SCHIST, X35.

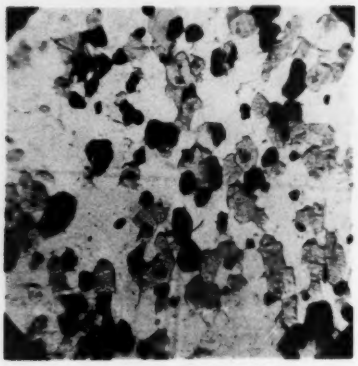

FIGURE 4. FROM KEWEENAWAN BASALT, X30.

PHOTOMICROGRAPHS OF HORNFELSES IN DULUTH GABBRO

NOTE 4. RELATION OF LAMPROPHYRES TO INCLUSIONS

Beger has noted (1923) that the "differentiation tendency" of lamprophyres is analogous to that of "basisches Ausscheidungen", as he calls certain inclusions. Several other writers have noted certain similarities in texture or composition in inclusions and lamprophyres (Daly, 1912-1913, p. 349) (Pabst, 1928, p. 362) (Goodspeed, 1930) (Jeremine, 1931) (Pl. 4). The writer has seen at Saganaga Lake a greenstone inclusion that appeared to have been partly melted and graded into a lamprophyre dike, which cuts its enclosing granite and includes granite fragments.[24] The relationship is highly confusing, but it seems as if both host and inclusion must have had a "mushy" consistency, such that they could break or glide, and yet flow into the fracture, or inject the glide planes.

It may be suggested that inclusions and lamprophyres, related to large stocks and batholiths, differ according to the time they are included. Early inclusions may be completely assimilated or disintegrated and mixed into the host; later ones may be so altered as to form unrecognizable inclusions or lamprophyres; and very late ones remain recognizable xenoliths. There may even be a sequence of effects in the lamprophyre stage; early lamprophyres become broken and confusedly mingled with the host; later ones give the fragmented dikes (List V) (Pl. 6) and very late ones remain dikes with more connected forms.

As a speculation, it may be suggested that, if most inclusions are altered fragments of country rock and if most lamprophyres are half-melted fragments of country rock, they should be expected to be petrographically similar. In spite of the confusion involved in such a relationship, it is believed to be a more logical explanation than the suggestion that granites have late siliceous residues, which form aplite and pegmatite at the same time that late basic residues form lamprophyres. And, again, it is difficult to assume that some basic differentiates form dikes of lamprophyre and others segregate into inclusions a few inches across. Both these suggestions may be retained as working hypotheses, if anyone finds them of interest, but it is believed that little evidence will be found to favor them. More and more, geologists seem to find explanation of the late basic magma in the softening of basic wall rocks and fragments. If such material is half melted, it can furnish, by hybridism and contaminaton, all the complexity found in the troublesome lamprophyre group. This in no way modifies the evidence that some lamprophyre dikes are in sediments far from intrusive stocks. The peculiar features of lamprophyres may very well originate in more ways than one.

MINOR NOTES

Note 5.—The field association of large rock masses with inclusions of the same character in the intrusive is of great value and almost universally used if the association is discovered (List I, 1, List II, 1, List III, 1, and List IV, 2). Examples can be cited, however, to show that, in many places, inclusions from one wall rock are carried by magma into the vicinity of different wall rocks (Powers, 1915) (William-

[24] F. F. Grout: *Origin of the igneous rocks of Minnesota*, Jour. Geol., vol. 41 (1933) p. 216.

son, 1935). If the character of the inclusions is different from that of any rock cropping out in the district, it may still be possible that contact action could make such inclusions as are found, from only one of the known rocks, and the association is still of some value (List III, 3). The recognition of the source rock in doubtful cases may depend on some petrographic peculiarity, such as an odd feldspar (Hurlbut, 1935), an orthorhombic pyroxene, an intergrowth (Grout, 1926), amygdules (Macgregor, 1931), and other local features. The lack of any such rock as the inclusions, and such as could produce the inclusions by contact action, in the larger outcrops in a district, is strong evidence that the inclusions had a different source.

Note 6.—The association (or lack of association) of inclusions of uncertain character with others of known origin has not led to any certain conclusions (List I, 3). The history of a batholith or stock is long and complex; the early inclusions may be much more changed than later contributions, and many of them may travel far into strange associations.

Note 7.—A series of inclusions showing all gradations from a doubtful mass to a recognizable fragment of some known associated formation is useful (Ghosh, 1934) (List I, 3b and 6, List II, 2, List III, 3). A quantitative chemical or mineralogical study may show that mineral ratios form a series (Nockolds, 1935-1937), even if the series was not clear in the field. If the series leads at one extreme to the host rock containing the inclusion, it has been supposed to indicate that the inclusion segregated (Daly, 1903b, p. 43) (List VI, 12), but this is wholly unsafe as a conclusion. Reaction and strew (page 1530) produce gradations from almost any xenolith to its host (Wright and Wright, 1908). If the intrusive is differentiated into a series from light to dark facies and the inclusions in each facies are a little darker than the host rock (List VI, 8), it is considered to be a sign of segregation (Pabst, 1928); but reaction might explain a progressive change in the color tone of inclusions with increasing feldspar as the surrounding magma evolved (Bowen, 1922, p. 541, 553, 559).

Note 8.—The sizes of inclusions are noted in five of the lists of criteria. There is such a divergence of opinion that no value can be put on the criterion unless it is a local feature.

Note 9.—The shapes are of little value. It has been recognized since 1880, or earlier, that concretions may be broken like fragments, and fragments may be rounded like concretions (Phillips, 1880), but sharp angularity may be enough to indicate rather strongly that an inclusion was not segregated in place.

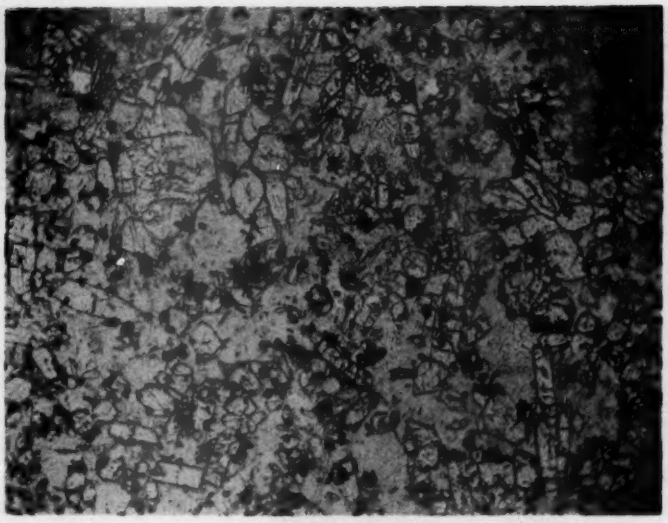

FIGURE 1. AUGITE LAMPROPHYRE AT SAGANAGA LAKE, X50.

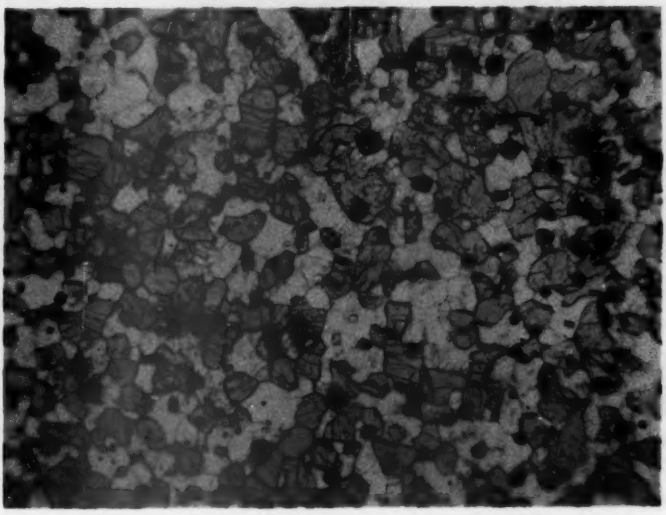

FIGURE 2. HORNFELS FROM GUNFLINT IRON-BEARING FORMATION, X35.

RESEMBLANCE BETWEEN LAMPROPHYRE AND HORNFELS TEXTURES

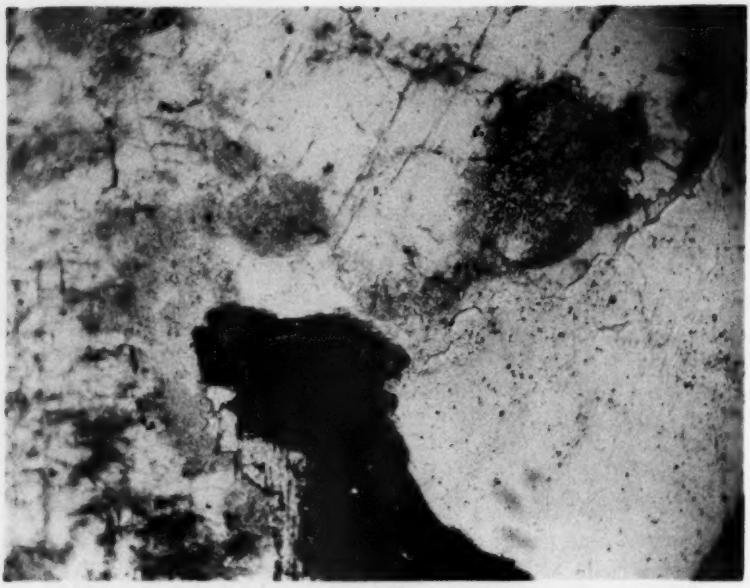

FIGURE 1. THIN SECTION OF GIANTS RANGE GRANITE, X80

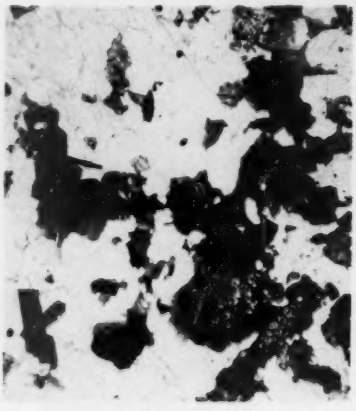

FIGURE 2. GIANTS RANGE GRANITE
Recrystallized to a hornfels by Duluth gabbro, X30.

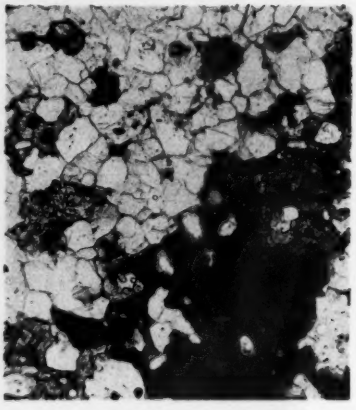

FIGURE 3. HORNFELS FROM OGISHKE-KNIFE LAKE SEDIMENTS

INCLUSIONS IN DULUTH GABBRO

Note change of texture of granite by contact action, and similar poikilitic textures in two hornfelses from different original rocks. X55.

Note 10.—The sharpness of contacts of inclusions is often noted, but sharp contacts may appear in any sort of inclusion (unless possibly in one segregated in place, and such are not well known). On the other hand, any sort of inclusion may show interlocking across the contact, and gradations (Bastin, 1908) (Pl. 11, fig. 1).

Note 11.—Bedding in inclusions is generally a clear indication of sedimentary xenoliths, and may be preserved through metamorphism (Hall and Gardthausen, 1911) (List I, 4). Its absence, however, is no sign that the inclusions are not sedimentary, because reaction may obscure the bedding (Grout, 1933).

Note 12.—Schistose and gneissic structures may be in inclusions of any origin, and where a fragment is oriented by the flow of the host magma the schistosity may, or may not, be in harmony with the flow structure of the host. Perhaps more than half the schistose inclusions may be xenoliths, as Phillips thought (1880), but they may be of either igneous or sedimentary origin; and many intrusions have a primary foliation so that cognate fragments are gneissic. Any sort of inclusion in a magma that has flow structures may be softened so that it participates in the deformation (Adams and Barlow, 1910).

Note 13.—Several minerals have been listed as signs of sedimentary origin (List I, 7). Perhaps cordierite has been most discussed, and it will serve as an example. It has been described as crystallized from magma (probably contaminated by some sediment in solution) at Pallet (Lacroix, 1898); in the contaminated norite of the Bushveld Complex[25]; and in the contaminated granite at Cape Town[26]; it was described as a norite mineral in a rock from Minnesota,[27] but the rock was probably a hornfels inclusion. A reaction of magma and an older gneiss at Glen Shee is said to have resulted in biotite and magmatic cordierite (Williamson, 1935). In spite of the occurrences in contaminated igneous rocks, therefore, cordierite in an inclusion seems to be strongly suggestive of sedimentary origin.

The origin of hypersthene by reaction was rather fully discussed by Bowen (1922). Most minerals of igneous rocks may not be regarded as proving igneous origin, because they may also be formed in inclusions by reaction. Possibly such a rare mineral as anorthite may not have been found except with direct igneous origin. The concentrations char-

[25] A. L. Hall: *The Bushveld igneous complex of the central Transvaal*, Geol. Surv. South Africa, Mem. 28 (1932) p. 312.

[26] A. L. DuToit: *The geology of South Africa*, London (1926) p. 132.

[27] A. N. Winchell: *Mineralogical and petrographic study of the gabbroid rocks of Minnesota; and more particularly of the plagioclasytes,* [illegible]. Geologist, vol. 26 (1900) parts of p. 151-388.

acteristic of anorthosite and peridotite are rare except in igneous rocks (Pl. 7).

Certain mineral peculiarities, such as color, or inclusions, or content of rare elements (Brammall and Harwood, 1932), may serve to identify the original rock that contained these minerals. The absence of such peculiarities, however, is not a sign that the inclusions came from another source, because reaction may change the minerals in these characters.

Note 14.—Biotite and hornblende are dominant mafic minerals in inclusions in the abundant granites. Statistically, the biotite schist inclusions are dominantly of sedimentary origin (List I, 8) and the hornblende schist inclusions are dominantly of basic igneous origin (List II, 4) (Grout, 1933, p. 1027). Reaction might be expected to turn all the mafic minerals in an intrusive and its inclusions to the same final product, but many granites have both biotite and hornblende, so that neither mineral in an inclusion is rapidly attacked.

Note 15.—Reactions between magma and inclusions are discussed in a masterly way by Bowen (1922), and with the addition of a few later discussions (Read, 1924) (Nockolds, 1933 and 1934) the student will find that paper sufficient.

Some xenoliths are so large that reaction and the mineral series from the cores to the outer zones has been incomplete; surrounding contaminated igneous rocks are then of the utmost value in interpretation (List I, 6; List II, 18; List III, 10; List IV, 11). Zones of different composition are to be expected in any sort of inclusion but are perhaps less likely in cognate inclusions than in xenoliths. Reaction can occur, however, not only with xenoliths but with early solidified parts of the intrusive itself. Light zones around inclusions are not definite criteria that the inclusion has segregated, because, in many examples, xenoliths of quartzite or schist have light zones [28] around them (Sederholm, 1926) (Jones, 1930) (Vogt, 1930). The magma zone near an inclusion may be dark from strew (p. 1530) or light because reaction leaves the liquid enriched in the light constituents toward the low temperature end of the reaction series (Bowen, 1922, p. 559-560).

Some orbicules are zoned, and others are not (Pl. 8); there has been a strong tendency to interpret some as xenoliths and others as segregations. So many have xenoliths in the center that perhaps all should be regarded as derived from xenoliths, some having reacted more completely than others.

[28] F. F. Grout: *Anorthosite and granite as differentiates of a diabase sill on Pigeon Point, Minnesota*, Geol. Soc. Am., Bull., vol. 39 (1928) p. 572-573.

FIGURE 1. MAIN INTRUSIVE SYENITE
Photo by Robert Balk.

FIGURE 2. APLITE WHICH FOLLOWED FRACTURE FIRST OCCUPIED BY LAMPROPHYRE IN CONTACT ROCKS

LAMPROPHYRE DIKES GRADING INTO INCLUSIONS AT SNOWBANK LAKE

FIGURE 1. OLIVINE MASSES IN BASALT

FIGURE 2. ANORTHOSITE FRAGMENTS IN DIABASE
Many are rounded masses or, if much broken, single crystals (xenocrysts). Larger masses may have segregated, but, in this exposure, blocks are clearly broken.

INCLUSIONS WITH MINERAL COMPOSITIONS INDICATING IGNEOUS ORIGIN

Note 16.—"True igneous textures" are often noted as evidence of igneous origin (List II, 8; List IV, 13), but these textures are not easily distinguished from some in recrystallized sediments. There may be in slate fragments, for example, a development like igneous porphyritic, poikilitic, and even granitoid textures, so that the criterion is of little value. The reverse criterion, however, is more important, and the value of the typical hornfels texture is discussed in Note 3. The retention of traces of ellipsoids, amygdules, glomeroporphyry, or other such primary igneous features by inclusions is perfectly satisfactory evidence of their igneous origin (List II, 5; List III, 5).

Note 17.—A hornblende-rich inclusion in a granite bearing euhedral hornblendes (List VI, 21) might be considered to be a sign of segregation from the magma (Harker, 1909, p. 348) (Pabst, 1928, p. 368). Bowen's study of reaction (1922), however, shows that if an inclusion has any dark mineral in abundance it will probably react to form those minerals that are also in the magma around. Furthermore, not all the dark minerals of inclusion are early; it is recorded (Smyth and Buddington, 1926, p. 79) that a "segregation" contained late magnetite and titanite. Evidently, the criterion is wholly unsafe.

Note 18.—The common olivine masses in basalt (Pl. 7, fig. 1) may result from accumulation of crystals on a shelf or floor, with later violent intrusive motion to break off fragments (Pirsson, 1926) (Harker, 1909, p. 346); or may be jammed by motion of a magma through a constriction, and broken up by injection.[29] Such an origin for inclusions (List IV, 5) is probable in volcanic outlets but is hardly likely in the broad chambers of stocks and batholiths.

Note 19.—Gradation from dikes to fragments is clearly seen in some exposures (Fig. 1 and Pl. 6). It is not always clear whether the dike formed before the main intrusive was solid (Grout, 1929) (Balk and Grout, 1934), after solid and before palingenically softened (Sederholm, 1923, 1926, 1934), or after solid and at depths where rocks flow in the solid state. Similar structures in the Grenville sediments in Canada are attributed to rock flow rather than to palingenesis.[30]

SUMMARY COMMENT

The critical notes make clear that several of the suggested criteria are not sound. They also emphasize the fact that there is no good

[29] N. L. Bowen: *Crystallization-differentiation in igneous magmas*, Jour. Geol., vol. 27 (1919) p. 425.

[30] D. H. Newland: *Plastic deformation of Grenville limestone*, N. Y. State Mus., Bull. 196 (1917) p. 145-147. In an older report, Ebenezer Emmons once suggested that the limestone was injected as a plutonic [*Geology of New York, Pt. II, comprising the survey of the second geological district* (1842) p. 228-229].

example of an inclusion that originated by segregation in place, except for those special varieties, the glomeroporphyries (Harker, 1909, p. 346) (Bailey, 1924, p. 290) and possibly the orbicules (Lahee, 1931) (Pl. 8 and Pl. 9, fig. 1). Many have suggested that a continued segregation like that forming a glomeroporphyry might form an anorthosite inclusion. Inclusions of anorthosite can be formed in other ways (Pl. 7, fig. 2), however, and the proof that any inclusion larger than a glomerophenocryst has formed by segregation has not yet been offered. The anorthosite masses with glomerophenocrysts may be accidental associates, brought together by gravity; there are, at some places, xenoliths of still other rocks associated with glomerophenocrysts.

Some of the zoned orbicules may be segregated,[31] but these are so rare and so different (Pl. 8) from common inclusions that the suggestion that the common inclusions originated by the same process is not well supported (Johannsen, 1932).

Segregated lumps, a few inches across, are reported in scores of districts, but the evidences of origin are woefully weak. Nearly always the origin by segregation is assumed wholly from the similarity of minerals in the inclusion and the matrix. As Nockolds says (1932, p. 451), "The number of examples where *xenoliths* (also including under this term many of the 'basic segregations' of older writers) of basic igneous rock, enclosed in granite, *bear no relation to the surrounding rocks is gradually increasing.*" (Italics are the writer's.)

Pabst (1928) and Hurlbut (1935) write as if the process of liquation was distinct from that of segregation. It is a process by which segregation may have occurred, and might explain the nearly spherical forms that are common, but it is as doubtful as other processes as an explanation for dark inclusions in granitic rocks. Liquation has been abandoned as an hypothesis by most of those who suggested it. It remains to be asked by what other processes can segregation take place forming dark inclusions.

The term cumulophyric[32] implies that minerals become aggregated in forming clusters in much the same way as the glomerophenocrysts.[33] The term glomeroplasmatic[34] implies that minerals may be localized in a magma "in open clusters", aggregates of one or several minerals, in a granular mass of different composition; the rock for which the term was suggested, however, has hypersthene and a granulitic texture characteristic of reaction of a shale with magma.

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[31] J. J. Sederholm: *On orbicular granites*, Comm. Geol. de Finland, Bull., no. 83 (1928) p. 1-105. This gives 98 references.

[32] Whitman Cross, J. P. Iddings, L. V. Pirsson, and H. S. Washington: *The texture of igneous rocks*, Jour. Geol., vol. 14 (1906) p. 703.

[33] Alfred Harker (1909, p. 347) pictures "glomeroporphyry" with an aggregate of several minerals.

[34] Loewinson-Lessing: Soc. Imp. Nat. St. Petersburg, Tr., vol. 30 (1900) p. 208.

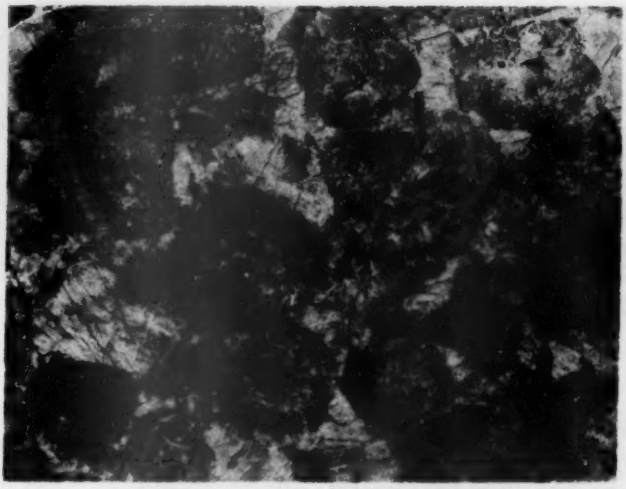


FIGURE 1. NO RECOGNIZABLE XENOLITHS
May have resulted from segregation. From Norway. X .5.

FIGURE 2. MICA SCHIST CORES IN ORBICULES
From Idaho. X .5.

ORBICULAR GRANITES

FIGURE 1. GLOMEROPORPHYRY, COOK COUNTY, MINNESOTA
About natural size.

FIGURE 2. MOTTLED SHONKINITE, BORDER OF VERMILION GRANITE
Scale shown by folded ruler one foot long. Such rocks readily identified in some cognate xenoliths.

FIGURE 3. GREENSTONE BRECCIA
Swarm from borders of Giants Range granite.

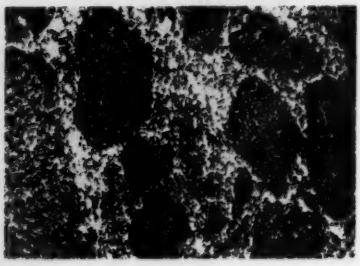


FIGURE 4. GREENSTONE BRECCIA
Greenstone fragments show differences in rounding and in reaction with Giants Range magma. About one-half natural size.

GLOMEROPORPHYRY, SHONKINITE, AND BRECCIA

INCLUSIONS IN CERTAIN SIGNIFICANT MASSES

GENERAL STATEMENT

The writer's own field studies of inclusions have led him to a progressive change of opinion as to the origin of inclusions; beginning with the Duluth gabbro of Minnesota, in which a variety of rocks develop a hornfels texture; later, the Boulder batholith of Montana, in which the inclusions have "phenocrysts" that look igneous; the Vermilion granite of Minnesota, with mica schist from slate and cognate fragments of shonkinite; the Giants Range batholith of Minnesota, inclusions of which contain added "phenocrysts"; the Alta stock of Utah, which has inclusions of a porphyritic early-cooled, border facies; and the Saganaga granite of Minnesota, which appears to have been intruded where no sediments were available, so that the inclusions are mostly from basalt flows. The clear evidence in these intrusives, together with some experience in about as many others, emphasizes the lack of value in most of the criteria that have been used but leaves a residue of a few points, which, when found, can be considered highly suggestive.

INCLUSIONS IN THE DULUTH GABBRO

The inclusions in the Duluth gabbro [35] are largely derived from slate, but furnish a particularly good example of the way a magma changes the mineral and chemical composition of slate inclusions to make them resemble igneous rocks. Several other kinds of inclusions and contact effects have some bearing on the criteria here discussed. An early anorthosite facies supplies cognate fragments to a late granitic facies (List IV, 1, 2 and 4).

The Duluth gabbro has in its floor several formations, chiefly slates, but some graywackes, conglomerates, iron-bearing formations, basalts, and a variety of older intrusives, diabase, syenite, and granite. Contact metamorphism extends for nearly half a mile. The common result is a recrystallization to hornfels, even the granites and syenites being made "sugary-grained" for a few feet (Allison, 1925, p. 496-7) (Pls. 2, 3, 4, 5).

The inclusions in the gabbro are most numerous near the floor, and practically all the formations known in the floor have been recognized as inclusions.[36] Sugary-grained hornfels makes probably 99 per cent of all the inclusions, and those near the slate floor are largely slate hornfelses (List I, 1), but detailed petrographic features serve to identify others as derived from certain other originals. The thin-bedded slates

[35] Data are taken largely from earlier papers by the writer: *Probable extent of abyssal assimilation*, Geol. Soc. Am., Bull., vol. 41 (1930) p. 678-684; *Contact metamorphism of the slates of Minnesota by granite and by gabbro magmas*, vol. 44 (1933) p. 1008-1033.

[36] One report of granite included in the gabbro seems to be contact-altered conglomerate in the floor, not really included.

form thinly banded hornfels at places (List I, 4b), but the slaty cleavage has disappeared, and the banding seems to have resulted from differences in beds (Pl. 2, fig. 3). Most of the slate hornfels has lost the banding, and there are gradations (List I, 6). Then there are basalt flows turned to hornfelses almost identical with those from slate (Pl. 3, fig. 4), but at Duluth the hornfels has locally some altered amygdules (List II, 5) (List III, 5) and more olivine than the common slate hornfels. The graywackes and conglomerates have become hornfels that differs little from that derived from slate except that it has more coarse poikilitic biotite (List I, 8). The diabase intrusives of the gabbro floor are recrystallized to hornfels that is distinguished from others only in those local masses that bear large labradorite phenocrysts (Pl. 3, fig. 2) (List II, 6). The iron-bearing formations also yield some characteristic hornfels texture and a mixture of stable minerals including quartz (Pl. 3, fig. 1) (List I, 2, and 4a).

Associated with the rocks that by contact action converge to a hornfels are a few that are notably different. The iron-bearing rocks high in silica, or high in iron oxide, recrystallize where included in gabbro near by (List I, 1) to form coarse quartz or magnetite rocks respectively (List I, 2). There are also mixed facies of the formation that by recrystallization are turned almost wholly to one silicate mineral, amphibole or pyroxene; and the resulting texture is coarse rather than sugary, like a hornfels. A few sandstone beds become quartzites.

More than two-thirds of the material in the floor of the gabbro is slate, so that probably more than two-thirds of the inclusions of hornfels in the gabbro are of sedimentary origin (List I, 1 and 10).

The original slate minerals are chiefly quartz, a variety of feldspars, sericite, chlorite, and biotite. The slate hornfels is chiefly labradorite, pyroxene, biotite, and cordierite, with locally quartz, magnetite, hornblende, garnet, epidote, muscovite, and others. A few inclusions in the gabbro where thoroughly recrystallized carry olivine. The pyroxene is monoclinic in some and orthorhombic in others, both kinds being together in a few. Most of the included hornfels has plagioclase clouded with dusty black inclusions (List II, 16), but the feldspar grew in place. This is especially true in the centers of the large grains, which are much coarser than the grains in the original sediment.

Chemical changes in the slate hornfels are indicated in Table 3. Silica, water, and potash are lower in the hornfels than in the slate; but iron, magnesia, lime, and titanium are higher (Tables 1 and 2). The change clearly makes the slate hornfels resemble a gabbro in composition, so that it has lost all commonly recognized criteria of sedimentary origin.

Probably at Duluth there was little exchange of material between the

TABLE 3.—*Analyses of slates, and fragments of the same, included in the Duluth gabbro, Minnesota*

	UNALTERED FORMATION					ALTERED FORMATION				
	Knife Lake				Animikie	Knife Lake slate inclusions			Animikie slate inclusions	
	Type locality		South of Tower		Drill core					
	Slate	Graywacke	Slate	Graywacke	Slate					
	1	2	3	4	5	6	7	8	9	10
SiO_2	54.71	61.39	63.88	61.44	58.04	50.16	47.45	52.84	49.07	51.01
Al_2O_3	20.52	16.97	17.70	15.53	18.66	15.70	20.92	23.62	17.21	16.12
Fe_2O_3	1.72	.39	3.02	.97	1.51	1.83	2.87	.65	.46	.62
FeO	6.40	5.32	1.80	4.20	5.98	4.98	6.24	10.00	12.68	9.08
MgO	4.76	3.84	3.72	3.74	3.24	9.04	5.70	3.16	3.60	7.62
CaO	1.93	3.21	2.72	7.06	1.02	15.96	9.85	3.92	9.66	10.05
Na_2O	2.83	2.78	1.78	1.99	2.12	1.52	2.14	2.64	2.96	2.44
K_2O	2.68	1.25	3.34	2.13	3.28	.02	2.89	.67	trace	.35
H_2O+	3.25	2.44	.87	1.08	3.28	.24	.92	1.87	1.55	.78
H_2O-	.15	.06	.23	.08	.38	None	None	.....	.....	.09
CO_2	.....	.88	.....	1.73	.14	.....	.....	.....	2.70	.89
TiO_2	.89	.62	.52	.53	.71	.70	1.42	trace	.....	.69
ZrO_2	.....	.07	trace	.....	.13	.....	.....	.....	.....	.....
P_2O_5	.....	.19	.58	.....	.12	.....	.29	trace	.....	.07
SO_3	.....	.....	.....	.....	.70	.....	.....	.....	.....	.....
S	.....	.15	.19	.....	.17	.....	.....	.....	.....	.....
Cr_2O_3	.....	.01	.....	.....	None	.....	.....	.....	.....	.....
MnO	.....	.12	.....	.....	.08	.....	.02	.43	trace	.16
BaO	.....	.06	.26	.....	.44	.....	.....	.....	.....	.....
Organic matter	.....	.....	.....	.....	.56	.....	.....	.....	.....	.....
Totals	99.84	99.75	100.61	100.48	100.56	100.15	100.71	99.80	99.89	99.97
Specific gravity	2.81	.....	2.68	.....	.....	.....	.....	2.83+	.....	.....

1. Knife Lake slate at Knife Lake, Grout and Hartwell, analysts.

2. Knife Lake graywacke at Knife Lake, F. Grout, analyst.

3. Knife Lake slate, south of Tower, D. Manuel, analyst.

4. Knife Lake graywacke, south of Tower, Yarwood and Grout, analysts.

5. Virginia slate drill core, 88 to 1723 feet, on Mesabi range, Sec. 1, T. 57 N., R. 17 W., Alworth Hole. George Ward, analyst.

6. Hornfels inclusion in Duluth gabbro near Frazer Lake, where the main wall rock is Knife Lake slate. Yarwood and Grout, analysts.

7. Hornfels from Knife Lake slate included in gabbro at Jordan Lake. W. S. Yarwood, analyst.

8. Probably hornfels, "cordierite-norite." Sec. 15, T. 63 N., R. 9 W., near Snowbank Lake. Included in gabbro where the main wall rock is Knife Lake slate. A. N. Winchell, analyst [Am. Geologist, vol. 26, p. 303].

9. Hornfels inclusion of large size, with some bedding, near Bashitanaqueb Lake, Sec. 2, T. 64 N., R. 5 W., where the main wall rock is Animikie slate. A. D. Meeds, analyst [Minn. Geol. Nat. Hist. Survey, 21st Ann. Rept., p. 150-151].

10. Hornfels inclusion near the base of the gabbro. Island in Gabemichigama. The inclusion shows bedding and the associated iron formation inclusions suggest that all are derived from the Animikie, although some Knife Lake slate is near by. R. B. Ellestad, analyst.

included slate and the magma. Gabbro emanations changed the contact rocks almost to the composition of a gabbro before they were included (Grout, 1933, p. 1014-1016). Some of the larger blocks, however, may have been included when only slightly altered, and may have affected the magma by reaction. The reaction effects of the rarer iron-bearing rocks and quartzites are recognized and indicate that the gabbro was quite capable of reaction. Quartz gabbros occur near quartzite [37] and chert [38] inclusions, and iron oxide forms notable ilmenite schlieren near inclusions of Gunflint iron-bearing rocks. The formations were originally so different from bed to bed that the amount of change cannot be tested.

It is clear that, in the Duluth gabbro, inclusions of some compositions retain their sedimentary minerals and chemical features, but that the common slates, graywackes, and basic igneous rocks lose their identity, becoming hornfelses that show their origin only in a few rare cases by retaining some petrographic peculiarity such as bedding, amygdules, or phenocrysts.

Segregated titaniferous iron ores are coarse bands in the gabbro rather than lumps or block-like inclusions. The coarse texture (List VI, 1) contrasts with the hornfels texture in most of the inclusions.

INCLUSIONS IN THE VERMILION GRANITE

The main part of the Vermilion batholith is a pink biotite granite [39] with petrographic characters close to those of hundreds of other biotite granites. Partial analyses of seven samples show a fair degree of uniformity over an area 30 by 80 miles. The walls are chiefly slates of the Knife Lake formation, just referred to in discussing the Duluth gabbro; but there are places where the wall rock is greenstone, and others where it is older granite gneiss. The contact metamorphic zone is from 100 yards to 10 miles wide. Basic border facies of the granite are exposed locally along both the north and the south sides and grade through hornblende granite, hornblende syenite, to shonkinite and hornblendite. As the greenstone is altered at the contacts to hornblende rocks, it might be suspected that the border facies resulted from contamination; but the same rocks appear at the border where older gneiss constitutes the wall, and there they cannot be attributed to contamination. One of the petrographically peculiar border facies is a mottled shonkinite (Pl. 9, fig. 2).

[37] W. S. Bayley: *The eruptive and sedimentary rocks of Pigeon Point, Minnesota, and their contact phenomena*, U. S. Geol. Surv., Bull. 109 (1893).

[38] T. M. Broderick: *The relation of the titaniferous magnetites of northeastern Minnesota to the Duluth gabbro*, Econ. Geol., vol. 12 (1917) p. 663-696.

[39] Data are taken largely from Grout, 1926.

Thousands of inclusions in the granite are of biotite schist, which the best criteria (List I, 1, 2, 4, 5, 7, and 8) clearly show are derived from the slates which become such schists near the granite. Inclusions range in size up to masses so many miles long that they seem to be pendants of the roof rather than detached blocks. Except where the schist is injected lit-par-lit the composition is so much like that of the slate that not much reaction can be detected (Grout, 1933, p. 1004-1005). At one place, a schist inclusion is bordered by a magnetite concentration. At another, the biotite granite near a group of schist inclusions seems darker than normal, and chemical tests suggest that about 20 per cent of biotite schist may have been assimilated by the magma before final consolidation (Pl. 1, fig. 2) (List I, 21).

A much smaller number of inclusions are hornblende schist, probably derived from greenstone that makes up part of the nearby wall rock (List II, 1 and 4). The hornblendic inclusions furnish examples of hybrids and of strewing about of hornblende to form gneisses and ghost-like traces of reaction inclusions (List II, 2) (List III, 3).[40]

At one instructive place the granite contains cognate inclusions of the spotted shonkinite, which is the peculiar basic border facies of the granite (List IV, 1 and 2). This is north of Pine Island, on the north side of Vermilion Lake. Because the spotted shonkinite is so odd in petrographic character and because its origin is so definitely determined by its local occurrence along the borders of the granite regardless of the wall rock, inclusions of the rock in granite are excellent examples of cognate fragments.

INCLUSIONS IN THE BOULDER BATHOLITH

The relation of the Boulder batholith to the ore deposits of Butte, Helena, and other districts has justified a series of studies of the batholith itself. The inclusions in the batholith have been noted and discussed without much agreement (Weed, 1912) (Winchell, 1914) (Billingsley, 1915-1916) (Billingsley and Grimes, 1917-1918) (Stansfield, 1928) (Tansley, Schafer, and Hart, 1933).

The main intrusive is a quartz monzonite, with an early diorite facies at the north and south ends and scattered aplitic granites nearer the center. The wall rocks range from a wide variety of sediments and andesites to Archean schists and gneisses. Contact metamorphism has whitened the gray limestone for a considerable distance but has turned it to garnet and epidote for only a few feet. It has made a hornfels of the andesite for a few feet at some places, but seems to have had little effect on the Archean schist and gneiss. A series of steeply dipping

[40] G. M. Schwartz: *The contrast in the effect of granite and gabbro intrusions on the Ely greenstone*, Jour. Geol., vol. 32 (1924) p. 108.

sediments on Big Hole River are largely shales, and these also have been recrystallized to hornfels near the batholith.[41]

Inclusions, a few inches across, of darker color and finer grain than the main intrusive are sparsely scattered in most outcrops (Pl. 1, fig. 1). Locally, they are larger and more abundant (Pl. 10, and Pl. 11, fig. 1) and constitute "swarms" in which the bulk of the dark inclusions is greater than that of the matrix. The best known of these areas is that from Dewey down the Big Hole River gorge for a few miles, at the southwest corner of the exposures of the mass. These have been attributed to cognate xenoliths, fragments of early border facies, basic secretions, igneous xenoliths, and foreign fragments of other rocks.

South of the power dam between Divide and Dewey the auto road cuts into the bluff to make an excellent exposure of the contact, and north of the same dam the old railway-cut gives a second exposure nearly as good. On the railway, the proportions of material in the sedimentary series can be estimated closely; shales are largely dominant, with only a few limy beds and greenish cherty-looking layers making 5 or 10 per cent of the section, alternating with the shale. On the road, it can be seen that just at the contact the shale has been brecciated by intrusion so as to resemble a conglomerate; but this breccia follows the intrusive contact rather than the sedimentary bedding, and there can be no doubt it is a result of intrusion. The hornfels merges gradually to shale, within perhaps 100 yards from the intrusive. Inside the intrusive the inclusions nearest the hornfels resemble the hornfels in nearly every detail (List I, 2 and 10). Inclusions are so numerous, however, that among them there are many with differing characters; and at greater distances from the contact the inclusions have a more igneous appearance (List I, 3b and 6). These igneous-looking lumps are the objects of disagreement in past reports. Feldspar "phenocrysts" are abundant in many, and the composition and minerals are so close to those of the quartz monzonite matrix that the inclusions might well be considered of igneous origin. It is noteworthy, also, that a few inclusions of limestone and of cherty-looking sediment are still recognizable without any alteration to hornfels, in the midst of the dominant "porphyritic" inclusions. It is argued that, because these known sedimentary inclusions are not changed, the igneous-looking inclusions are not altered sediments.

There are several doubtful points in the argument. The scattered chert inclusions show that the batholith was actively incorporating some of its sedimentary roof and walls. A study of the district around indi-

[41] F. F. Grout and Robert Balk: *Internal structures in the Boulder batholith*, Geol. Soc. Am., Bull., vol. 45 (1934) p. 877-895.

FIGURE 1. ELONGATED SWARM CROSSED BY PEGMATITES

FIGURE 2. SWARM IN WHICH MOST INCLUSIONS SEEM STRETCHED OR ELONGATED WITH MOTION OF MAGMA AROUND

DARK INCLUSIONS IN BOULDER BATHOLITH

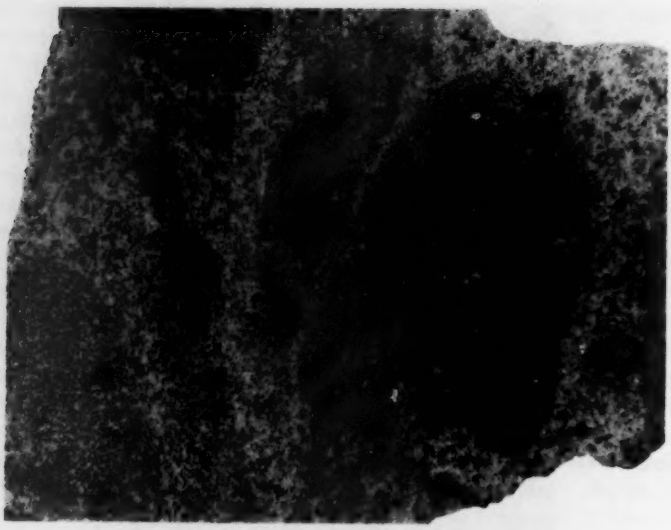


FIGURE 1. DARK INCLUSIONS IN BOULDER BATHOLITH
Some distinctly outlined; others grade into schlieren between those that are distinct. X .5.

FIGURE 2. GRANODIORITE FROM ALTA STOCK
Dark-gray inclusion, whose length is parallel to hornblende needles; both are normal to aplite dikes. X .5.

cates that there is much more shale than chert available, so that there should be about ten times as many fragments of shale as chert among the inclusions. The only fragments as abundant as that, however, are the igneous-looking inclusions, so that it is evident the shale inclusions have changed so as to look igneous (List I, 3a). It must also be added that, whereas the early border facies of the intrusive has about the composition of some of the dark inclusions, no border facies of porphyritic texture has been recorded. Probably more than half the inclusions here are "porphyritic," and many of the rest are finer than the matrix.

Some geologists might question whether or not the intrusive was capable of changing a shale inclusion into one with such igneous characters. This is the same problem that arises the world over, and this exposure seemed to be a good one for careful study. The writer selected samples from the sediments near the Boulder batholith, from the contact rocks, from inclusions near the contact, and from inclusions progressively farther inside the batholith.

The shale hornfels outside the brecciated contact zone has a grain size about 0.04 mm. and contains about 40 per cent quartz, 25 per cent feldspar, 30 per cent biotite and other mafic minerals, and a small percentage of several minor accessories. The texture is granoblastic, "sugary," and analogous to shale hornfels elsewhere.

The hornfels of the contact breccia shows a grain size about 0.07 mm. and contains 40 per cent quartz, 50 per cent biotite, chlorite, and other mafic minerals, 6 per cent muscovite and sericite, and 2 per cent magnetite, with accessory pyrite, apatite, and leucoxene.

An inclusion with exactly the appearance of the hornfels contact rock was found six inches inside the contact in the intrusive. It has a grain size about 0.17 mm. and contains 6 per cent quartz, 40 per cent labradorite, 10 per cent biotite, 35 per cent amphibole, 5 per cent magnetite, and a little pyrite. The amphibole seems to have formed at the expense of biotite and quartz.

An inclusion in the great swarm along the banks of the Big Hole River, about 100 yards from the contact, had a porphyritic texture with "phenocrysts", 1 to 3 mm. long, in a matrix that had an average grain of 0.3 mm. This contained 2 per cent quartz, 60 per cent labradorite, 20 per cent hornblende, 15 per cent biotite, and 3 per cent magnetite.

This petrographic series certainly seems to show that rocks near the contact are intermediate between those that are clearly sedimentary beds and those that look wholly igneous but occur only as inclusions so far as known (List I, 3b); all are probably of sedimentary origin.

Analyses are added for the sake of being quantitative in a second method of study. Table 4 shows that, although the formation is not exactly uniform and the changes in series are not regularly progressive, there is a transition from shale to porphyry. With the added data of

Table 5, it is clear that the inclusions of "porphyry" contain more lime, magnesia, iron oxide, and titania, and less potash, silica, and water than the shales. These are exactly the changes known to occur elsewhere in series from slate to included hornfels (Tables 1 and 2). It

TABLE 4.—*Gradation in alkalies and iron oxide from Cretaceous shale to basic inclusions, in and near the Boulder batholith*

	Total iron as Fe_2O_3	Na_2O	K_2O	Na_2O^* $+K_2O$
1. Shale 100 yards outside monzonite	5.62	.51	2.36	2.87
2. Shale 1–100 yards composite	4.66	1.28	2.36	3.64
3. Shale about 20 yards outside	4.58	1.77	2.69	4.46
4. Hornfels breccia at contact (1931)	6.01	1.71	1.60	3.31
5. Hornfels breccia at contact (1932)	4.34	1.41	1.74	3.15
(contact)				
6. Hornfels inclusion 6 inches inside intrusive	8.87	.94	1.10	2.04
7. Sedimentary hornfels inclusion near contact	3.10	.61	.70	1.31
8. Inclusion 100 yards inside intrusive	9.46	3.32	1.53	4.85
9. Composite of inclusions 1 to 500 yards in from contact	10.23	2.82	1.55	4.37
10. Typical inclusion far in the monzonite	10.31	3.76	1.45	5.21
11. Composite of many inclusions near Homestake near center of batholith	9.28	4.18	2.08	6.26

* Determinations by R. B. Ellestad and T. Kameda.

seems entirely possible, as far as minerals, texture, and composition are concerned, that the inclusions are derived from shale. The argument that other included sediments are not so greatly changed simply means that shale is more promptly recrystallized into a different-looking rock than are limestones or cherty-looking beds.[42] It must also be noted that this great swarm of inclusions which has aroused so much discussion occurs, not everywhere about the batholith, but exactly where shale wall-rocks are prominently exposed (List I, 1). The strongest evidence against a cognate igneous origin for the inclusions is the lack of any such rocks as early facies of the batholith. The early basic facies are diorites and, if contaminated, hornblendites,[43] but they are not like the "porphyry" inclusions.

Table 5 gives more complete analyses of some of the shale and of the inclusions that seem to be derived from it, by reaction with the magma. The igneous character of the analyses of inclusions is apparent, but cannot be considered a criterion of igneous origin.

[42] "Cherty-looking" is used, rather than chert, for these rocks on Big Hole River because two were tested, one from the sediments outside and one from an inclusion; they had each between 60 per cent and 70 per cent silica.

[43] Verner Jones: *Spring Hill gold deposit, near Helena, Montana*, Econ. Geol., vol. 29 (1934) p. 544-559.

The difference between this occurrence and that at Duluth lies chiefly in the fact that the few plagioclase phenocrysts in inclusions at Duluth are large and basic (Pl. 3, fig. 2) and indicate that the original was a porphyritic diabase such as occurs nearby (List I, 2), whereas at the Big Hole River the plagioclase "phenocrysts" are smaller, more

TABLE 5.—*Analyses of shale, hornfels, and inclusions in monzonite of the Boulder batholith along the Big Hole River east of Dewey, Montana*

Constituent	Shale	Shale Hornfels	Hornfels Breccia	Hornfels inclusions	Dark inclusions	Dark inclusions
	1	3	4	6	8	10
SiO_2	65.53	69.73	67.33	55.15	51.97	49.18
Al_2O_3	16.82	15.95	14.57	17.95	19.38	20.24
Fe_2O_3	.38	.49	.74	.37	3.71	3.72
FeO	4.72	3.68	4.74	7.65	5.18	5.93
MgO	2.44	1.76	2.67	5.71	3.45	4.62
CaO	4.00	1.61	2.79	8.16	8.22	8.69
Na_2O	.51	1.77	1.71	.94	3.32	3.76
K_2O	2.36	2.69	1.60	1.10	1.53	1.45
H_2O+	1.32	.99	1.96	1.01	.73	.72
H_2O-	.25	.26	.22	.17	.11	.05
CO_2	.18	.05	.24	.11	.16	.....
TiO_2	.71	.61	.63	.96	1.29	.74
ZrO_2	.....	.02	.01	.....	none	.....
P_2O_5	.24	.24	.27	.24	.31	.44
S	.06	.05	.32	.....	.03	.....
MnO	.10	.06	.04	.24	.16	.17
BaO	.07	.11	.04	.....	.05	.....
Cl	.....	.....	.....	.....	.02	.....
	99.69	100.07	99.88	99.76	99.62	99.71

1. Cretaceous shale, 100 yards outside the monzonite. T. Kameda, analyst.
3. Cretaceous shale hornfels, near contact of monzonite. T. Kameda, analyst.
4. Hornfels breccia, from shale, at contact of monzonite. T. Kameda, analyst.
6. Hornfels inclusion in monzonite, six inches from contact with shale hornfels. T. Kameda, analyst.
8. Dark inclusion in monzonite, 100 yards from contact. T. Kameda, analyst.
10. Typical dark inclusion in monzonite, 300 yards from contact. R. B. Ellestad, analyst.

silicic, and such as might be expected to grow in the hornfels by reaction. This is a long step to take in reasoning and might be left uncertain except for the broader field relations—lack of such porphyry in the wall rock of the Boulder batholith and the presence of early basic facies that are not like the inclusions. It is freely admitted that recrystallization might turn a coarse-grained diorite to a fine-grained hornfels, as noted by Joplin (1935) and as already recorded as an effect of Duluth gabbro on granite and syenite (p. 1555). None of these, however, produces a characteristic porphyry appearance. (*See* the later paragraphs on Alta and Giants Range.)

INCLUSIONS IN THE ALTA STOCK, UTAH

The grandiorite intrusive at Alta[44] is of about as much interest, from the standpoint of ore deposits as the Boulder batholith, and it includes along its borders large fragments from its sedimentary walls. Much of the wall rock is limestone, and, at the contact, it is garnetized and mineralized. Besides these, small dark inclusions (Pl. 11, fig. 2), mostly only a few inches across and mostly carrying phenocrysts of feldspar, are widely scattered in the mass. These at places make up swarms a few feet wide and perhaps 50 or 100 feet long. The writer saw no swarm in which inclusions make up 50 per cent of the belt in which they are concentrated.

The whole stock was studied, and numerous observations were made concerning the inclusions. There is no evident relation between the abundance of the inclusions and the occurrence of shale or hornfels in the adjoining wall rocks. There was found, however, near the southwest side of the mass, an outcrop of dark-gray porphyry, much like the inclusions, which may well have been an early facies of the intrusive (List II, 1, 6 and List IV, 1). The boundary of the whole mass would seem to enclose a single intrusive unit if the porphyry is mapped as part of the main intrusive, but would have a peculiar re-entrant if the porphyry were excluded. The area of porphyry is much too large to be considered an inclusion of the same sort as those a few inches across, scattered elsewhere through the mass.

In petrographic character, these porphyry inclusions are almost identical with those of the Boulder batholith, and might have been assumed to be altered shale xenoliths if it had not been for the discovery of the intrusive porphyry *en masse* at Alta. This discovery changed the interpretation completely, and it seems likely that the porphyry inclusions here represent cognate fragments of igneous origin.

As a speculation, it may be suggested that, if there had been a little more vigorous injection of the main grandiorite, later injection might have entirely shattered the early bordering porphyry, turning it all into inclusions. In the absence of any outcrop of porphyry, it would have been easy to mistake these porphyry fragments for shale xenoliths, with metacrysts. Possibly the inclusions in the Boulder batholith at Big Hole River may have resulted from such vigorous injection of an early porphyry, now lost; but the present evidence is against that idea, because only one criterion favors it and a group of five criteria point the other way.

[44] B. S. Butler, G. F. Loughlin, V. C. Heikes, et al.: *The ore deposits of Utah*, U. S. Geol. Surv., Prof. Paper 111 (1920) pl. XXVII.

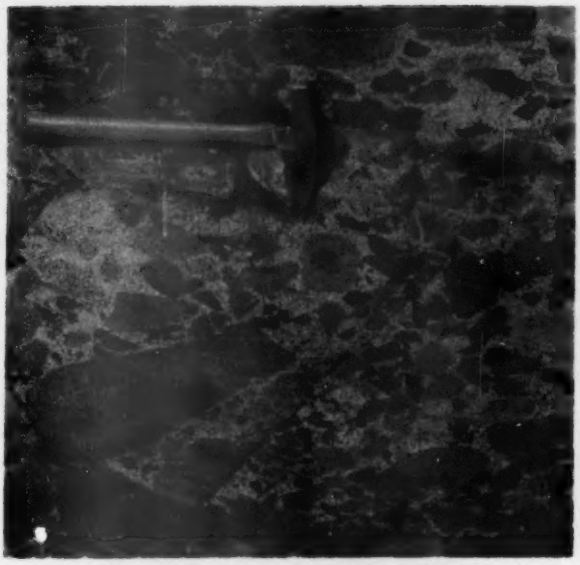


FIGURE 1. SWARM OF GREENSTONE FRAGMENTS
Differ in recrystallization and apparently in rounding by attack of magma.

FIGURE 2. FISH-SHAPED GREENSTONE INCLUSIONS PARALLEL TO LIGHT SCHLIEREN

INCLUSIONS IN SAGANAGA GRANITE

FIGURE 1. SLIGHTLY ALTERED, X50.

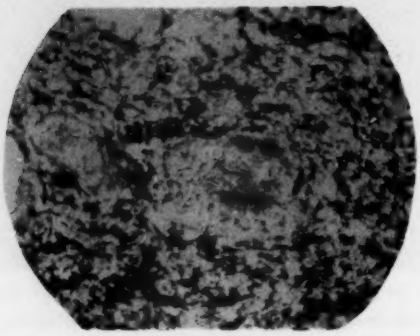


FIGURE 2. FELDSPAR REPLACING SCHIST BUT INCLUDING ORIENTED DARK NEEDLES (HELECITIC), X25.

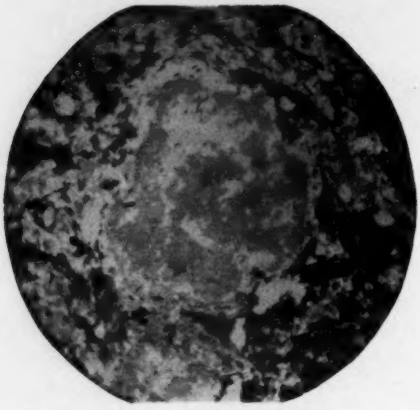


FIGURE 3. FELDSPAR CROWDING ASIDE NEEDLES, DISPLACING RATHER THAN REPLACING THEM (OCELLAR), X12.

PHOTOMICROGRAPHS OF BIOTITE SCHIST IN GIANTS RANGE GRANITE

INCLUSIONS IN THE SAGANAGA GRANITE

The Saganaga granite has been described previously as to petrography (Grout, 1929) and structure.[45] It is a light sodic hornblende granite of Laurentian age, older than the sediments of the district, and intrusive only into greenstone (Archean basalts) and older gneisses.

The inclusions in the granite are mostly greenstone, more or less recrystallized to hornblende schist and hornblendite (Pl. 12) (List II, 1, 2 and 4; and List III, 1 and 3). There are few doubtful "phenocrysts" in the inclusions. There are some bodies of uncertain character, however, even here. First, near the greenstone walls where greenstone inclusions are numerous, there are some inclusions of sugary or hornfels texture and, in the same exposures, sugary lamprophyre dikes that seem in some way related—possibly the inclusions were heated to plastic masses which could move into fractures (List V, 1). A dike fragmented in a complex way, forming inclusions, is the basis of the sketch shown in Figure 1. Second, there are many dark patches of various sizes, farther from the greenstone walls, some sharply bounded and others more gradational. These dark facies at several places show a peculiar petrographic consanguinity with the granite around them—namely, they have coarse quartz aggregates or "eyes," which do not occur in most of the rocks of the Lake Superior region but do occur almost everywhere in the Saganaga mass. If the dark masses are altered greenstone xenoliths, they had quartz and probably feldspar introduced into them at a late magmatic stage, when quartz was deuterically corroding the early minerals of the granite. This is comparable with additions to igneous xenoliths at Tregastel (Thomas and Smith, 1932). It is possible, however, that they are lamprophyric differentiates or hybrids with an even closer relation to the granite. Nearly all the dark rocks of the district are either greenstones or lamprophyres. Near one of the large lamprophyres there is an apparent gradation from dark rocks to granite as if by contamination rather than by segregation. The border facies, only slightly more basic than the main mass of the granite, is not commonly so dark as these. The evidence here is inconclusive as to the origin of dark patches with quartz "eyes".

INCLUSIONS IN THE GIANTS RANGE GRANITE

The Giants Range batholith intrudes the pre-Cambrian complex along the north side of the Mesabi iron range of Minnesota for 100 miles. It is partly hornblende granite and partly a later biotite granite, partly uniform-grained and partly spotted with large feldspar phenocrysts. Contact metamorphism affects a wide zone on each side, turning green-

[45] F. F. Grout: *Structural features of the Saganaga granite of Minnesota-Ontario*, 16th Inter. Geol. Congr. 1933, Rept., vol. 1 (1936) p. 255-270.

stones to hornblende schist, and slates to mica schists. These two schists are the chief formations that could supply xenoliths to contaminate the magma (Allison, 1925).

Inclusions have been noted at several places, but most of them consist of hornblende schist or biotite schist and furnish no special problems. Toward the east end of the batholith, there are exposures of the breccia at the contact of the greenstone (Pl. 9) (List II, 1, 2, 4) (List III, 1, 3), and of the granite recrystallized by the contact action of the later gabbro (Pl. 5, figs. 1 and 2); the grain is reduced, from about a millimeter in width, to a mosaic or poikilitic state, with many grains a fifth of a millimeter wide. Two occurrences are noteworthy.

The biotite granite northwest of Mountain Iron in section 31, T. 59 N., R. 18 W. has large-sized cognate fragments of the earlier hornblendic and somewhat gneissoid facies of the batholith (List IV, 1). Followed to the east, these fragments are increasingly more numerous and larger up to the main outcrops of gneissoid hornblendic granite (List IV, 2). There can be little doubt of their cognate origin.

North of Virginia, the biotite schist inclusions in hornblende granite near the biotite-schist wall rock are instructive. Many inclusions resemble the wall rock (Pl. 14, fig. 1) but are associated with others that look like a dark porphyry (Pl. 14, figs. 2 and 3). The same problem arises as in the Boulder and the Alta intrusives. Here in the Giants Range the penetration of the granite into the schist, transforming it into a rock of wholly igneous appearance, was evident in field exposures (Pl. 15). One can hardly question the injection of tiny apophyses of granite, the smallest of which grade down to the size of single crystals of feldspar; and the most instructive inclusions show a few phenocrysts beyond apophyses for a short distance only. Farther from the main contact, large inclusions of "porphyry" have uniformly scattered "phenocrysts" (Pl. 14, fig. 3), but it seems that they, or at least certain constituents which resulted in their growth, have been added to a mica schist of sedimentary origin. No such porphyry is known in the region as a border facies or older intrusive, except at places where the granite could have added the "phenocrysts". Plate 13, figures 2, 3, show that some added feldspars grow by replacement (List I, 9), but probably other metacrysts displace, rather than replace, the schist.

This outcrop seemed important enough to deserve further check by chemical analysis to determine the nature of the change when "phenocrysts" were added. Petrographically the biotite schist is much more hornblendic and much less schistose where the "phenocrysts" appear in the inclusions in hornblende granite, and it is possible that the hornblendic facies is not derived from the adjacent slate wall, but from a

hornblendic greenstone to which feldspar "phenocrysts" were added. All the inclusions studied, however, come from outcrops very close to mica slate walls, and this fact with the evident gradational series strongly indicates progressive additions from the magma.[46] Table 6 contrasts the original biotite schist of sedimentary origin at the contact with a

TABLE 6.—*Analyses of mica schist contact rock and "porphyry" inclusion in the Giants Range of Minnesota*

	1	2
SiO_2	64.29	53.56
Al_2O_3	16.05	15.64
Fe_2O_3	.69	1.88
FeO	4.58	5.76
MgO	3.25	6.91
CaO	2.58	7.14
Na_2O	3.09	4.21
K_2O	2.53	1.59
H_2O+	1.44	1.57
H_2O-	.04	.08
CO_2	p. n. d.	p. n. d.
TiO_2	.56	.87
P_2O_5	.22	.34
S	.....	.04
MnO	.06	.13
BaO	.....	.08
	99.38	99.80

1. Biotite schist at south wall of Giants Range batholith. Sec. 28, T. 59 N., R. 17 W., Minnesota. R. T. Biscoe, student analyst. The percents of soda and potash have been checked.
2. Inclusion with "phenocrysts." Same location. Duncan McConnell, student analyst.

"porphyry" inclusion near by, included in the hornblende granite where it was evident that the granite could add coarse feldspars (Pl. 15). The additions and losses are much like those in other districts (Tables 1 and 2).

INCLUSIONS IN THE SIERRA NEVADA

The writer has not made a field study of the material of the Sierra Nevada, but a long series of papers (Turner, 1894) (Knopf and Thelen, 1904-1906) (Gilbert, 1906) (Knopf, 1918) (Pabst, 1928) (Cloos, 1931) (Miller, 1931) (Nockolds, 1932) (Mayo, 1935) (Hurlbut, 1935) have contributed data, so that there may be some value in a critical review of the findings. Similar criticism might be undertaken for several districts.

[46] If the original rock was greenstone the details would be modified, but the conclusions stand: the change in most constituents would be less striking, but additions of lime and soda to form feldspars are clearly indicated.

The Sierra Nevada batholith is one of the largest, and is complex in its petrographic and structural relations. The major facies may be close to granodiorite. Its dark inclusions are of several kinds. There are clearly sedimentary xenoliths and many other the origin of which is not clear—they have been called basic segregations, autoliths, clots, reaction inclusions, and inclusions of older diorite or gabbro. The texture of inclusions is commonly finer than that of the intrusive, and many are "porphyritic."

The most elaborate discussion of these bodies is by Pabst (1928), who calls them "autoliths" in spite of admitted uncertainty as to their origin. He realizes that no early facies of the batholith resembles the inclusions, so that he hesitates to call them cognate fragments, but he seems to favor the idea of segregation. The inclusions are so perfectly analogous to those already described as reaction xenoliths in the Giants Range and Boulder batholiths that the writer believes there is little sign of segregation. On the other hand, Nockolds, with a different experience on other intrusives (and equally little personal knowledge of the Sierras), says (1932) of the inclusions described by Pabst: "there seems little doubt that his 'autoliths' represent xenoliths of basic igneous rocks brought up from below . . . quite obviously igneous." Now, as the result of experiences in this country, the writer questions the value of those criteria that led Nockolds to consider the rocks "obviously igneous"—namely, the items in List II, under the subhead "not reliable criteria"; 7, uniform distribution and size; 8, "phenocrysts"; 13, fine grain; and 15, the dark color due to abundance of early minerals of the host. If there are no amygdules or similarly characteristic features of surface flows, and no outcrops of igneous rocks of analogous kinds in the district, it is wholly unsafe to consider inclusions igneous on the basis of such criteria. They are commonly derived from shale, by reaction and addition.

The somewhat similar inclusions studied by Miller (1931) and by Hurlbut (1935) in the southern Sierras seem to be derived from the petrographically similar early basic intrusive rocks that crop out near by (List III, 1 and 2). If they are, they furnish an example of reduction of grain size by inclusion and recrystallization, for they are finer grained than the gabbro or diorite of the older rocks. They also illustrate a difficulty in the use of the term "cognate," and in the criteria for determining that an inclusion is cognate. The older gabbros and diorites of the southern Sierras may be the immediate forerunners of the main batholith and somewhat related in origin. Does that somewhat remote relation make the fragments cognate? Is a rock derived from the same deep magma chamber in the same geologic period definitely cognate? By

FIGURE 1. COMMON DISTRIBUTION NEAR WALL ROCK

FIGURE 2. INCLUSIONS
Closer view of some with white metacrysts.

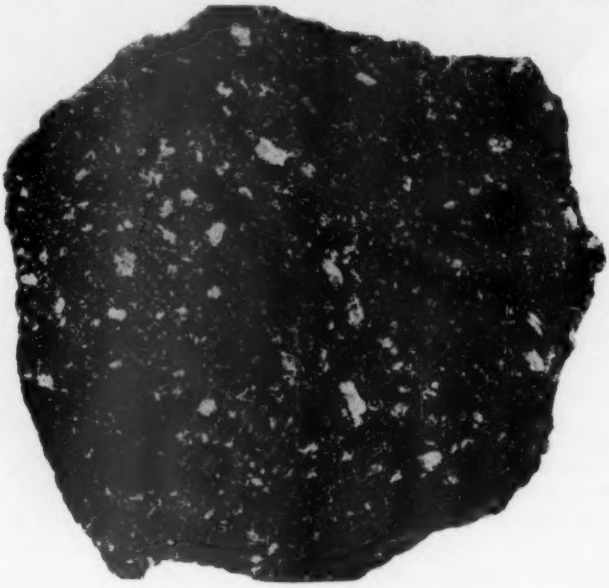
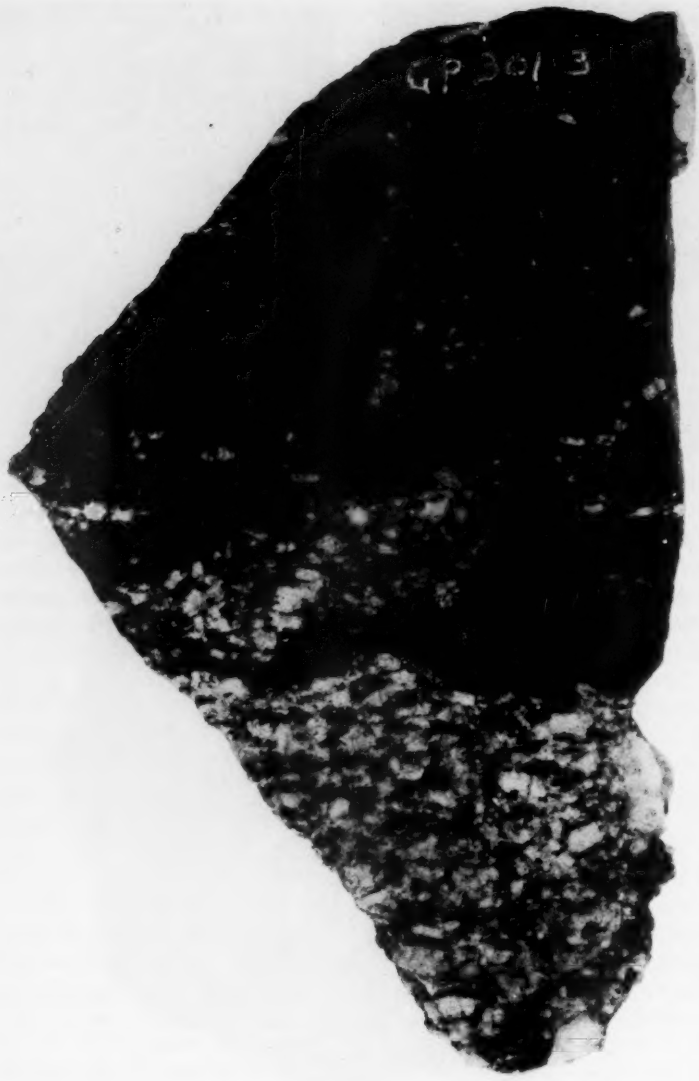


FIGURE 3. INCLUSION
With metacrysts.

SCHIST INCLUSIONS IN GIANTS RANGE GRANITE



CONTACT OF GIANTS RANGE GRANITE AND A SCHIST INCLUSION

Granite clearly worked its way into schist and added metacrysts beyond the irregular apophysis. About natural size.

the definition on page 1529, they would be excluded, they are vexing problems.

There are said to be other dark spots in the Sierra Nevada mass, consisting chiefly of hornblende, with the same grain as the hornblende in the surrounding intrusive. These are local and may be of different origin from the main troublesome inclusions.[47]

SUMMARY OF POINTS IN THE MASSES DESCRIBED

There are definitely recognizable cognate fragments in some batholiths, early facies solidified and included in later facies.

Shales commonly lose their sedimentary character and acquire several igneous features when included in magmas.

The sugary hornfels texture of an inclusion (or of the groundmass of an inclusion) generally means that it has recrystallized from a solid original, rather than having been formed by primary crystallization in a magma.

Shale hornfels xenoliths may acquire coarse feldspar crystals in a fine-grained groundmass, either by direct addition or by exchange reaction with the magma.

Not all porphyritic-looking inclusions, however, are derived from shale by reaction; there are early porphyritic facies, at least in stocks, which furnish some cognate fragments.

Shales are more rapidly and extensively recrystallized to hornfels textures than are igneous rocks or other sedimentary and metamorphic rocks.

The idea of exchange of material between a xenolith and a magma is a valid one, and exchange may have occurred in several of the masses, although the evidence of it is not conspicuous even at Duluth, where the rocks were most carefully tested; but a one-way action of magma on solid rocks is also a valid idea, and is proved by contact action of the gabbro on its wall rocks before they become included where they can react on the magma.

CONCLUDING REMARKS

The criteria used in estimating the origin of inclusions in large intrusives have been listed and critically discussed. A series of examples has been added to show that certain ideas are untenable and that others may be of value. The points that seem most in need of emphasis and attention are here stated briefly in final summary.

1. Segregation approximately in place has little support as an explanation for any inclusions except glomeroporphyries and possibly orbicules. The processes of segregation of small masses, so often suggested, are

[47] E. B. Mayo: *personal communication.*

questionable, and the certainty that most inclusions can be otherwise explained makes it inadvisable to call any inclusion a segregation unless the evidence is exceptionally strong. A heavy burden of proof rests upon anyone who claims to have a segregated lump.

2. Inclusions react with the magmas in which they are immersed. If their minerals are not such as are in equilibrium with the magma or its juices, they may be completely changed in minerals and textures and largely changed in chemical composition. The constituents of feldspars and of other minerals may be added by the magma to the inclusions, and some added materials may result in textures almost typically porphyritic. The criteria of origin are wholly lost. It is not safe to assume a particular origin, either cognate, or even igneous, when the criteria are thus lost. The masses may well be called by the noncommital term "inclusions." The reactions to expect were reported by Bowen in clear detail, and geologists should realize the significance of such reactions well enough to refrain from assumptions of igneous or cognate origin when the criteria are not there.

3. The sugary hornfels texture (granulitic, granoblastic, mosaic, decussate) is a strong criterion of recrystallization in a solid by contact action. The texture is approached by aplites and lamprophyres but not by the more common igneous rocks and groundmasses. Through most of this paper the writer has tried to maintain the attitude of an impartial judge, but at this point it seems advisable to be emphatic and strongly advocate the use of this criterion. Who ever saw a segregation that was sugary grained, or one that was porphyritic with a sugary groundmass? There has never been a well-authenticated example. Most of the fine-grained basic inclusions are derived from alteration of clays, but some are from the recrystallization of various other rocks.

The similarity of textures of lamprophyres and hornfelses suggests that lamprophyres may be derived from inclusions altered one step farther than most inclusions. "Complementary dikes" need restudy with this in mind. It is doubtful whether dikes are diaschistic.

4. Shales may react with magmas so thoroughly, so rapidly, and so far that large masses are rather uniformly changed. If this happens to clay fragments associated with other fragments in a batholith magma, some of the other rocks may be little changed, perhaps a centimeter or an inch along the outer margin. This should cause no surprise. As Bowen indicated, if the minerals of the inclusions are the same as those growing from the magma, little change is likely.

5. Shales are the most abundant sediments, and such reaction inclusions as may form by the action of magmas on clays should be expected to be abundant in most batholiths.

6. Field relations of inclusions to wall rocks, and gradational series are of much value in determining the origin of inclusions.

7. Where sedimentary minerals, structures, or compositions are present, they prove sedimentary xenoliths; their absence proves nothing.

8. The shapes, sizes, sharpness of contacts, zoning, and ratios of new minerals of inclusions are of little value in estimation of origin.

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STRATIGRAPHY AND STRUCTURE IN THE UPPER CHAMPLAIN VALLEY

BY JOHN RODGERS

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INTRODUCTION

This paper gives the results of a detailed survey of the Paleozoic geology of the Whitehall quadrangle, New York-Vermont, west of the Taconic thrust. Reconnaissance work has been carried northward from this area through a large part of the adjoining Ticonderoga quadrangle. These are standard 15-minute quadrangles of the United States Geological Survey. The field work was done in the summers of 1935 and 1936.

This survey is the first attempt to use the divisions of the "Calciferous"

formation, as established by Brainerd and Seely,[1] for more than a local area. It is entirely through the use of these divisions that the results given in this paper were obtained.

The author wishes to acknowledge his indebtedness to Mr. C. A. Hartnagel, who suggested work in this area; to Professor C. M. Nevin, who supervised the writing of the report; to Dr. D. H. Newland, Dr. Rudolf Ruedemann, and Miss Winifred Goldring; and to Dr. E. O. Ulrich and Dr. Josiah Bridge.

STRATIGRAPHY

PREVIOUS WORK

The formations discussed in this paper are those which constitute the western sequence of Paleozoic rocks in this region; that is, the sequence which was not involved in the great thrust-plates of the Taconic orogeny but outcrops between these and the Adirondack Mountains. The author has spent most of his time on the lower part of the section; for his knowledge of the beds above the Chazy he is indebted to Dr. G. Marshall Kay.[2]

The first geological work in this region was that of the New York State Geological Survey, particularly that of Ebenezer Emmons,[3] between 1836 and 1842. Emmons established the stratigraphic sequence of the region. In 1861, Hitchcock[4] published the results of his work in the Champlain Valley. In 1890 and the years following, Brainerd and Seely (op. cit. et alia) published the results of their work on the dolomite and limestone formations. Their work showed that the thickness of the Calciferous and Chazy formations had been greatly underestimated and that these formations could be divided into units convenient for mapping the structure of the area. Considerable work has since been done on the Chazy,[5] but the only use of the divisions of the Calciferous, other than incidental and often erroneous assignments of isolated outcrops to them, was that by Ruedemann[6] on the Port Henry quadrangle. Here he indicated the distribution of these divisions in the small areas of Calciferous on the New York side of Lake Champlain.

In 1911, Ulrich and Cushing[7] described the Calciferous of the Mohawk

[1] E. Brainerd and H. M. Seely: *The Calciferous formation in the Champlain Valley*, Am. Mus. Nat. Hist., Bull., vol. 3 (1890) p. 1-23.

[2] G. M. Kay: *Stratigraphy of the Trenton Group*. Geol. Soc. Am., Bull., vol. 48 (1937) p. 233-302; personal communication.

[3] E. Emmons: N. Y. State Geol. Survey, Ann. Repts. (1837-1840); *Natural History of New York*, Part IV, *Geology*, pt. 2 (1842) 437 pages.

[4] Edward Hitchcock: *Geology of Vermont* (1861) 2 vols., 558 and 982 pages.

[5] P. E. Raymond: *The Crown Point section*, Bull. Am. Paleont., vol. 3 (1902) p. 267-310.

[6] J. F. Kemp and Rudolf Ruedemann: *Geology of the Elizabethtown and Port Henry quadrangles*, N. Y. State Mus., Bull. 138 (1911) p. 65-68.

[7] E. O. Ulrich and H. P. Cushing: *Age and relations of the Little Falls dolomite (Calciferous) of the Mohawk Valley*, N. Y. State Mus., Bull. 140 (1911) p. 97-140.

Valley, with an incidental reference to two sections in the Champlain Valley. They divided the Calciferous into the Beekmantown and the Little Falls formations. The contact was drawn, on the basis of the incomplete section at Ticonderoga, in the middle of Division B of Brainerd and Seely. This correlation has been quite generally followed, and it has been assumed that there were two major pre-Chazy invasions of the sea on the eastern slope of the Adirondacks; namely, the Potsdam-Little Falls and the "Beekmantown".[8]

DESCRIPTION OF FORMATIONS

Potsdam sandstone.—The formations described are the units which were used in mapping the area, and although their boundaries were drawn at faunal breaks wherever possible, their fundamental basis is lithologic.

The Potsdam sandstone was originally described by Emmons.[9] It forms the lowest unit of the New York Paleozoic section except for beds in the thrust-plates of the Taconic orogeny.

The base of the Potsdam in this area is a coarse sandstone approaching a conglomerate. In many places it is arkosic. It grades upward into a fairly massive, clean sandstone. The cement is silica and the grains do not weather in relief. Ripple-marks are common in New York. In Vermont the rock has been metamorphosed slightly, and is more properly a quartzite. The thickness of the whole formation approaches 400 feet, although close to the Adirondacks the thickness may be considerably less.

At Whitehall, as elsewhere around the Adirondacks, the upper beds of the Potsdam show a transition into the Little Falls through interbedded dolomite and sandstone. This transition facies has been called the Theresa formation. At Shoreham, in the Ticonderoga quadrangle, the Little Falls is immediately underlain by 50 feet of massive quartzite without transition; beneath this is 50 feet of interbedded dolomite and sandstone, and then the body of the Potsdam. If these beds are equivalent to the "Theresa" at Whitehall, they bear out a suggestion of Emmons [10] in 1840 that the upper Potsdam graded from sandstone in the Champlain region to calcareous sediments in the Saratoga region.

Fossils have been recorded from beds of Theresa facies at several places; at Whitehall these beds correlate with the basal Franconia (Ironton), while on the north side of Adirondacks they are slightly higher in the sequence. Thus the upper Potsdam, as here used, is Franconia; the remainder is presumably Dresbach.

[8] Winifred Goldring: *Handbook of paleontology for beginners and amateurs, Part 2, The Formations*, N. Y. State Mus., Handbook 10 (1931) p. 268, 271.
[9] E. Emmons: N. Y. State Geol. Survey, 2nd Ann. Rept. (1838) p. 214.
[10] E. Emmons: N. Y. State Geol. Survey, 4th Ann. Rept. (1840) p. 347.

Little Falls formation.—The name Little Falls was originally proposed by Clarke [11] for the pre-Tribes Hill Calciferous of the Mohawk Valley. Ulrich and Cushing extended it into the Champlain Valley to include all of Division A and part of Division B of Brainerd and Seely. There is, however, a significant unconformity 35 feet below the base of Division B, and this break seems to correspond with the break above the Little Falls in the Mohawk Valley.

Brainerd and Seely's description of Division A follows:

"Dark iron-grey magnesian limestone, usually in beds one or two feet in thickness, more or less silicious, in some beds even approaching a sandstone. Nodules of white quartz are frequently seen in the upper layers, and near the top large irregular masses of impure black chert, which, when the calcareous matter is dissolved out by long exposure, often appears fibrous or scoriaceous. Thickness 310 feet." [12]

Of this, the upper cherty dolomite and sandstone layers do not belong to the Little Falls formation. The dark, almost black, gritty dolomite, with quartz stringers and crystals (Little Falls or Lake George diamonds) and occasional lighter layers, is typical of the Little Falls of the Champlain Valley and of the Mohawk Valley. It is also characteristic of the Gatesburg of Pennsylvania and of the Copper Ridge of Virginia and Tennessee.

Fossils are very scarce in the Little Falls. On the basis of lithology and position it is correlated with the Copper Ridge-Conococheague sequence of the Appalachians; this is correlated with the Trempeleau of the Mississippi Valley.

Whitehall formation.—The beds of Division B together with the upper 35 feet of Division A form a stratigraphic unit. They are separated from Division C above and from the Little Falls below by marked unconformities. They are here named the Whitehall formation.

The type section is on Skene Mountain, which rises out of the village of Whitehall, New York, on the east. The rocks dip rather gently to the east. The base of the Whitehall formation is exposed not far below the brow of the mountain on the steep west face; the base of Division C is exposed in a hillock to the east just across the first road east of the mountain.

The description of Division B as given by Brainerd and Seely is:

"Dove-colored limestone, intermingled with light grey dolomite, in massive beds; sometimes for a thickness of twelve or fifteen feet no planes of stratification are discernible. In the lower beds, and in those just above the middle, the dolomite predominates; the middle and upper beds are nearly pure limestone; other beds show on their weathered surfaces, raised reticulated lines of grey dolomite. Thickness 295 feet." [13]

[11] J. M. Clarke: *Classification of New York series of Geologic formations*, N. Y. State Mus., Handbook 19, 1st ed. (1903) p. 8, 16.

[12] E. Brainerd and H. M. Seely: *Op. cit.*, p. 2.

[13] *Op. cit.*, p. 2.

The dolomite in the lower part is usually extremely cherty. The base of the Whitehall is made up of 35 feet of alternating sandstone and dolomite. The dolomite layers are like those above—light, clean, and cherty. The sandstone beds are 2 to 4 feet thick, coarse, and massive; but the cement is not all silica and the grains weather into marked relief. There is only rudimentary bedding in the sandstone layers.

A search for fossils in the rocks of this formation at Shoreham resulted in finding a few poor forms. Ulrich kindly examined them for the author, and recognized them as correlative with the Strites Pond beds of the Philipsburg, Quebec, sequence, with the Hagerstown limestone of Maryland, with the Chepultepec dolomite of the southern Appalachians, and with the Gasconade of the Ozark section. The limestones of this age seem to be uniformly pure and fine-grained and to lie in general toward the east side of the Appalachian trough, while their equivalents toward the west are the Chepultepec and Larke dolomites. At Whitehall and at Shoreham, both facies are present, the limestone in general forming the upper part of the formation; at Ticonderoga, now 6 miles west of Shoreham (the distance has been somewhat lessened by minor thrusting), the upper half of the Whitehall is missing, and there is but little limestone interbedded with the extremely cherty dolomite.

Unnamed formation (Upper Calciferous).—Brainerd and Seely describe these beds as follows:

"*Division C.* 1. Grey, thin-bedded, fine-grained, calciferous sandstone on the edges often weathering in fine lines, forty or fifty to the inch, and resembling close-grained wood. Weathered fragments are frequently riddled with small holes, called *Scolithus minutus* by Mr. Wing. 60 ft.

2. Magnesian limestone in thick beds, weathering drab. 100 ft.

3. Sandstones, sometimes pure and firm, but usually calciferous or dolomitic. 70 ft.

4. Magnesian limestone like No. 2, frequently containing patches of black chert. 120 ft.

Thickness of C 350 ft.

"*Division D.* 1. Blue limestone in beds one or two feet thick, breaking with a flinty fracture; often with considerable dolomitic matter intermixed, giving the weathered surface a rough, curdled appearance; becoming more and more interstratified with calciferous sandstone in thin layers, which frequently weathers to a friable, ocherous rotten-stone. 80 ft.

2. Drab and brown magnesian limestone, containing also toward the middle several beds of tough sandstone. 75 ft.

3. Sandy limestone in thin beds, weathering on the edges in horizontal ridges one or two inches apart, giving to the escarpments a peculiar banded appearance. A few thin beds of pure limestone are interstratified with the siliceous limestone. 120 ft.

4. Blue limestone in thin beds, separated from each other by very thin, tough, slaty layers, which protrude on the weathered edges in undulating lines. The limestone often appears to be a conglomerate, the small enclosed pebbles being somewhat angular and arenaceous. 100 ft.

Thickness of D 375 ft.

"*Division E.* Fine-grained magnesian limestone in beds one or two feet in thickness, weathering drab, yellowish, or brown. Occasionally pure limestone layers occur, which are fossiliferous, and rarely thin layers of slate.

Thickness 470 ft."[14]

[14] *Op. cit.*, p. 2-3.

The thickness of the Upper Calciferous is therefore about 1200 feet. The three divisions are easily distinguished throughout the area examined. Division C has yielded the Lecanospira fauna of the Roubidoux dolomite of the Ozark region.

A large fauna was described from the upper half of Division D and Division E by Whitfield,[15] and these beds were named the Cassin formation by Cushing.[16] This fauna is found in the Smithville formation in the Ozarks. There appears to be a basal sandstone at the base of D-3, the base of the Cassin formation; sufficient faunal analysis has not yet been made to determine the relative importance of this and the other breaks in the sequence. Until this work has been done, it seems best not to apply names to any divisions of this sequence nor to the sequence as a whole. It will be referred to as the Upper Calciferous.

The name Beekmantown was originally proposed by Clarke and Schuchert[17] for the entire Calciferous; but since it became evident that the Calciferous spanned the break between the Cambrian and the Ordovician (or the Ozarkian and the Canadian of Ulrich), the term has been restricted to the Ordovician (Canadian) part of the sequence.[18] But there has been no general agreement as to the position of this break in the Champlain sequence. Ulrich and Cushing considered the break to lie between the middle of Division B and the base of Division C in the section at Ticonderoga where the upper part of Division B is lacking. They considered that the missing beds were part of the overlying formation, whereas actually they form a unit with the underlying, but in any case they excluded the Whitehall beds at Ticonderoga from the Beekmantown. But the Whitehall is now correlated with beds considered as basal Ordovician (although Ulrich holds that they are upper Ozarkian), and hence if the term Beekmantown is to be used for the Ordovician Calciferous, in its use in the Champlain Valley it should be extended to cover the Whitehall formation. The name Beekmantown has been the source of much confusion for some time, and the author regrets having to add to it.

The sandstone at the base of Division C, and resting unconformably upon the Whitehall formation, is quite distinctive in lithology. The grains are in general smaller than in either the Potsdam or the basal Whitehall sandstones, and this sandstone is much better laminated. Cross-bedding is often very well developed. The cement is dolomitic and the grains weather into relief, giving the peculiar structure referred

[15] R. P. Whitfield: Several papers in Am. Mus. Nat. Hist., Bulls. 1-3 (1886-1890).

[16] H. P. Cushing: *Geology of the Northern Adirondack region.* N. Y. State Mus., Bull. 95 (1905) p. 362-364.

[17] J. M. Clarke and Charles Schuchert: *The nomenclature of the New York series of geological formations,* Science, n.s., vol. 10 (1899) p. 876-877.

[18] E. O. Ulrich and H. P. Cushing: *op. cit.*

to in Brainerd and Seely's description. Moreover these characteristics are constant through over 50 feet of beds. The three sandstones, the Potsdam, the basal Whitehall, and C-1, are each distinctive, and are very useful guide horizons in working out the structure, provided it is borne in mind that thin sandstones with the characteristics of C-1 are sometimes found above (never below) that horizon.

Except at Shoreham, the sandstone of C-3 is thin or absent.

The upper 200 feet of Division D constitute a limestone unit, the thickest such unit below the Chazy. It forms the lower third of the Cassin formation of Cushing and appears to have a basal sandstone in places, which is like C-1 in lithology but not in thickness. The cement, however, is calcareous.

Chazy formation.—Limestones of Chazy age are present at Shoreham (200 feet) and at Orwell (50 feet). They are not found at Ticonderoga or in Benson or to the south. Only the middle division of the type Chazy (Crown Point formation) is present in the upper Champlain Valley.[19]

The limestone is fairly pure, and typically heavy-ledged; but it is sandy and even dolomitic in places.

Black River formation.—Owing to the considerable metamorphism of most of the outcrops of the post-Calciferous rocks of this area, the lithology of the Black River cannot be adequately compared with that of the exposures either in the northern Champlain Valley or elsewhere. It is a pure limestone, often quite fossiliferous, and usually fairly thick-bedded.

At Crown Point, both the Lowville and the Chaumont formations are recognized. The author has not attempted to differentiate them in the folded rocks of the Whitehall quadrangle. Here the Black River averages about 50 feet thick.

Trenton formation.—For purposes of mapping, the Trenton was divided into a limestone portion and a shale portion. The limestone portion was mapped with the Black River limestone. These are merely facies, and as such can be compared with (not correlated with) the rocks of similar facies in the lower Mohawk Valley.

The limestone portion averages 100 feet in thickness, and consists of thin-bedded shaly limestone, carrying characteristic Trenton fossils. This limestone is highly sheared in many places; in others, the lower beds are noticeably heavier-bedded and less shaly. At Crown Point, the Amsterdam (Rockland), Larrabee (Hull), and Shoreham (lowermost Sherman Fall) formations have been recognized by Kay.[20]

[19] P. E. Raymond: *The Crown Point section*, Bull. Am. Paleont., vol. 3 (1902) p. 272.

[20] G. M. Kay: *Stratigraphy of the Trenton Group*, Geol. Soc. Am., Bull., vol. 48 (1937) p. 259, 262-265.

Overlying the limestone is at least 500 feet of shale. It is everywhere involved in minor thrust faults, being repeated by the slicing. Toward the Taconic thrust on the east, it is metamorphosed into a low-grade slate, and underlies large areas, as near Benson.

In a few places in the slate areas, there are outcrops of a quartzitic slate, derived from a sandy shale. These outcrops are disconnected and have no apparent plan. Presumably they lie at the top of the section and represent an upper sandy facies of the Trenton shale. They would thus correspond in facies to the Schenectady beds of the lower Mohawk Valley. This facies has not hitherto been recognized in the Champlain Valley.

STRUCTURE

STRUCTURAL PROVINCES

The structural provinces of the Champlain and upper Hudson valleys are: the Adirondack Mountains on the west, the great thrust-plates in front of the Green Mountains on the east, and a belt of disturbed Paleozoic rocks of the western sequence between. Figure 1 shows the boundaries of these provinces on the Whitehall quadrangle, and an adjoining strip of the Ticonderoga quadrangle.

The Adirondack province in this area is characterized by great high-angle faults; the entire eastern and southern periphery of the Adirondack Mountains is cut by such faults. The rocks of the province are mainly pre-Cambrian; these rocks extend into Vermont in Bald Mountain. The presence of pre-Cambrian rocks here was first recognized by Hitchcock.[21] The rocks are Grenville sediments dipping gently (locally steeply) east, intruded by sills of granite. The summit ledge of the mountain is a granite sill, with flow-planes dipping a few degrees to the east, and pronounced flow-lines pitching due east. To the north of Bald Mountain, a considerable area underlain by Paleozoic rocks with low and variable dips has been included in the Adirondack province because it is free from the disturbance that characterizes the intermediate province.

The eastern province is bounded on the west by the Taconic thrust. Detailed work by Larrabee[22] on the slates in the southeast corner of the Whitehall quadrangle and adjacent areas shows that this province is characterized by low-angle thrust-faults and isoclinal folding with no high-angle faults other than tear faults. The rocks of the province belong to the Taconic sequence of Paleozoic rocks of this region, and are chiefly slates.

The intermediate province is narrower in the Whitehall quadrangle than anywhere else in the Champlain or upper Hudson valleys. It is

[21] Edward Hitchcock: *Geology of Vermont*, vol. 1 (1861) p. 264.

[22] D. M. Larrabee: *Colored slate belt of Vermont and New York*, Geol. Soc. Am., pr. 1936 (1937) p. 86; personal communication.

characterized by considerable disturbance, including both high-angle faulting of the Adirondack type and thrust-faulting. It is underlain by the Paleozoic sequence described in the first part of this paper.

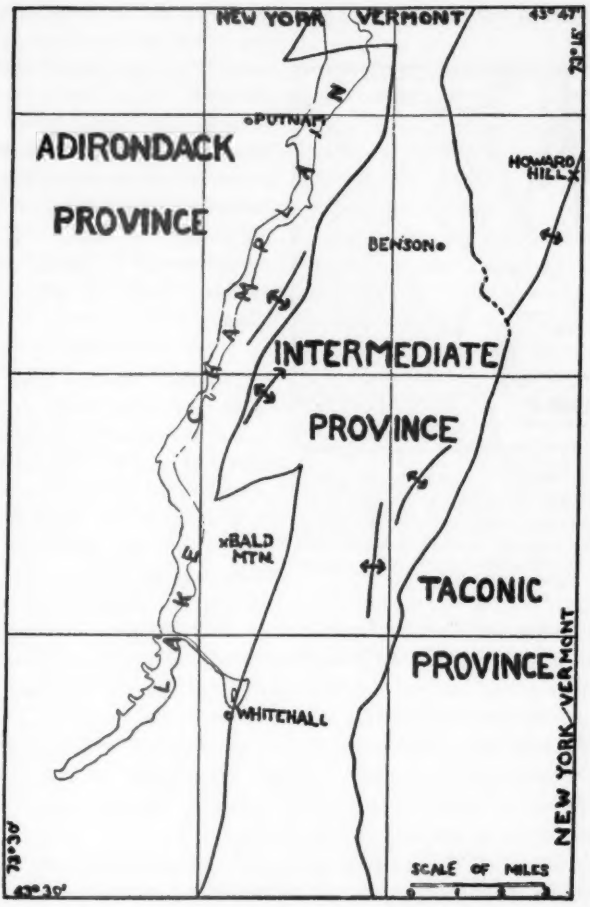

FIGURE 1.—*Map of the structural provinces of the Whitehall quadrangle*

FOLDS

Aside from the isoclinal folding of the Taconic province, the folds observed were open, simple, and remarkably scarce. The axes of the anticlines are plotted on Figure 1. The two western folds are not ac-

companied by synclines; the northern one is simply a low arch in the gently dipping Paleozoic rocks, the other is an elongate half-dome rather than an anticline. These folds are similar in trend and character—the writer believes them to be contemporaneous; they affect all the provinces and the one in the Taconic area warps the Taconic fault-surface, hence the writer believes them to be the last-formed structural feature of the area.

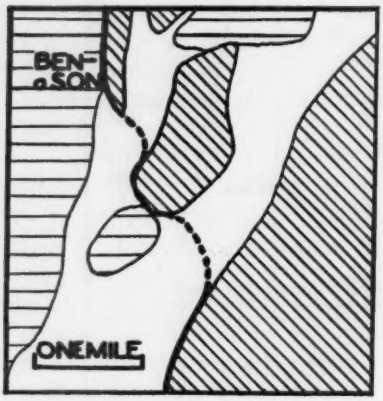

FIGURE 2.—*Sketch map of a small area at Benson and southwest of Howard Hill*

Areas of diagonal ruling, underlain by rocks of the Taconic sequence; areas of horizontal ruling, underlain by rocks of the western sequence; blank areas, covered by alluvium.

In a few places close to the trace of the Taconic thrust, the slates of the intermediate province were crumpled into practically isoclinal folds. These of course would be coeval with the isoclinal folds in the Taconic slates.

FAULTS

Taconic thrust.—This is the most prominent fault of the region. Its trace is shown on Figures 1, 3, and 4. It is marked by a mylonite zone in places.

The dip of the fault is low, as shown by the structure in the Howard Hill area. The fold shown on Figure 1 at Howard Hill so warps the Taconic fault-surface that limestone of the western sequence in the footwall of the fault outcrops on the top of Howard Hill at a height of 661 feet A. T., and slates of the Taconic sequence outcrop as low as 400 feet within a mile both northwest and southeast. Figure 2 shows the relations where this fold crosses the trace of the thrust. The fold is apparently double here. If the valleys east of the dotted line are underlain by slate of the western sequence, then the area of Taconic slate northwest of Howard Hill is a klippe.

But if the dip of the fault-surface is low, an interesting question arises in regard to its former extent. East of Whitehall, where it lies about 400 A. T., it is not more than 4 miles from pre-Cambrian summits that stand over 1000 feet A. T. Thus either the slates at one time overlapped the Adirondacks, or erosion since Ordovician time has been so slight as to cause the dip of the gently dipping fault-surface to migrate less than 4 miles.

High-angle faults.—The high-angle faults of the region can be grouped into four sets:

Longitudinal faults: Strikes ranging from N-S to N 40° E, averaging N 10° E;

Transverse faults: Strikes ranging from N 85° W to N 75° E, averaging N 80° E;

Northwest minor faults: Strikes approximately N 45° W;

Northeast minor faults: Strikes approximately N 45° E.

These faults are characterized by extensive brecciation, particularly when in dolomite; by well-developed trenches on the fault-lines; and by minor but conspicuous step-faulting. The erosion trenches range from 10 to 50 feet wide and from 5 to 40 feet deep, but the size of the valley is no measure of the importance of the fault. The step-faulting is well shown just north of Putnam where the main highway north climbs the hill; the lower ground to the east is underlain by the middle beds of the Potsdam sandstone; along the road at the base of the hill is a magnificent exposure of the contact between the Potsdam and the Grenville series; in the road-cut up the hill there is another higher exposure of the same contact; and finally the top of the hill is pre-Cambrian. A similar series of faults cuts off the north end of Bald Mountain.

The longitudinal faults dip (where determined) to the east at about 70 degrees. The transverse faults are probably nearly vertical. No attempt was made to determine the dip of the minor fault surfaces.

Figure 3 shows the pattern which these faults make in relation to the Taconic thrust. Where the faults are represented by solid lines, displacement was observed. Proved faults have been projected into the pre-Cambrian using the topography as a guide; they are then represented by dashed lines. The longitudinal and transverse faults cut the region into blocks, each block being about 5 miles north to south and 2 miles east to west. These show well on Figure 3 except in the northern part of the area where they are obscured by minor faults. The longitudinal faults lie parallel in general, and in places this parallelism extends to minor sinuosities. The pattern is strikingly similar to that found by Hudson [23] in the northern Champlain Valley.

Superimposed on this pattern are small faults parallel to the major longitudinal and transverse faults, and the two minor sets. The small faults of the longitudinal set always lie close to some major longitudinal fault; those of the transverse set are more irregularly distributed. The minor sets are prominent chiefly at the corners of the major blocks, as though they had been formed by shattering of these corners.

[23] G. H. Hudson: *The fault systems of the northern Champlain Valley*, N. Y. State Mus., Bull. 286 (1931) p. 5-59.

The downthrow of the longitudinal set is with one exception to the east. The fault next east of Lake Champlain and north of Bald Mountain varies from a stratigraphic throw of 1000 feet down to the east,

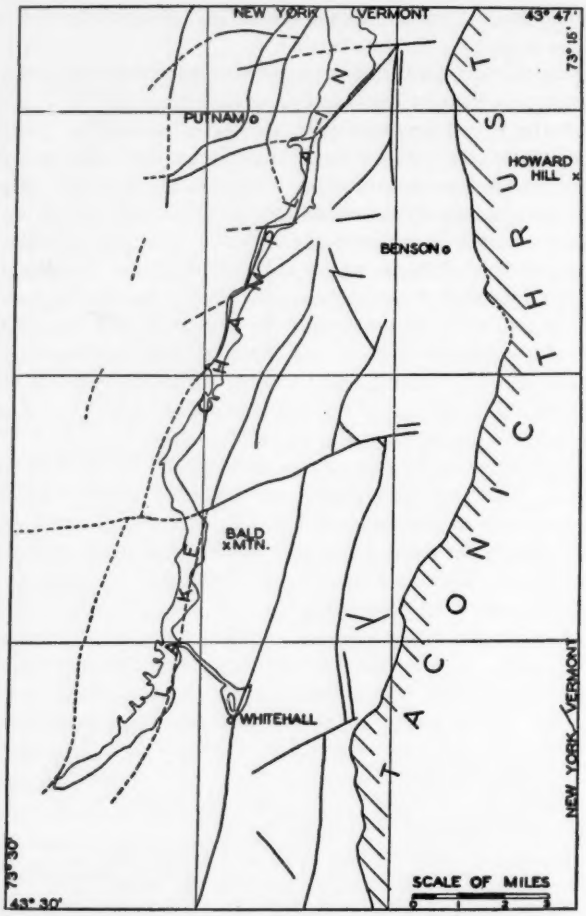

FIGURE 3.—*Map of high-angle faults of the Whitehall quadrangle*

directly west of Benson, to 1000 feet down to the west, only 4 miles to the south, where a small outlier of pre-Cambrian rock is brought up on the east side. The minor fault that parallels it through this distance has its throw consistently to the east. The stratigraphic throw of the

major fault is probably not its true throw, for this fault forms the boundary between the Adirondack and intermediate provinces, with their different structures. Hudson believes that movement took place on some of these longitudinal faults as early as Chazy time; if this is so, this boundary fault would presumably be affected, but the author has seen no evidence for motion before the deposition of the Trenton beds.

The major fault next east of the one just described has a throw of 2000 feet and cuts and follows closely a thrust fault of 1500 feet throw. Precisely the same relations hold for the easternmost major high-angle fault south of the transverse fault north of Bald Mountain; and the next fault west, the one east of Bald Mountain, forms the boundary between the Adirondack and the intermediate provinces. Thus there appears to have been over a mile of strike-slip on the transverse fault, the northern tier of blocks slipping to the west past the southern tier, which acted as a buttress. Strike-slip is the rule along the transverse faults, which displace all the other sets of high-angle faults in the area.

Yet with all this evidence of considerable throw along the high-angle faults, the eastern province is free from any comparable disturbance. The trace of the Taconic thrust shows no relation to or displacement by the system of faults to the west, and, as already mentioned, detailed work in the slates behind the thrust shows no high-angle faults except minor tear faults. The author concludes from this negative evidence that the movement of the Taconic thrust-plate into its present position took place after all movement had ceased along the high-angle faults.

Thrust faults other than the Taconic.—The dips of these faults are all to the east at angles of less than 45 degrees. They are normally parallel to the bedding or cleavage. They are marked by poorly defined crush-zones, and usually have no topographic expression other than that caused by the juxtaposition of two unequally resistant beds. This survey, though detailed, has probably discovered less than half of these thrust faults; particularly in those areas where the surface rock is shale there must be great numbers of these slices, mostly of small throw, which are difficult or impossible to detect.

Figure 4 shows the pattern which these faults make in relation to the Taconic thrust and the outcrop of the Adirondack pre-Cambrian. In the areas marked "shale" and "covered", there are probably as many of these faults as in the areas farther south. In the shale areas, the pattern has been filled in with symbols showing the generalized strike and dip of the cleavage. While some of the sinuosities are due to topography, the larger arcs cannot be accounted for in this way; they indicate overthrusting from the east. There seem to be three such arcs: one west of Benson, one south of Benson, and one east of Bald Mountain. Each

is approximately 5 miles from north to south, and each southern one overlaps its northern neighbor as though a later slice.

This progressive shift of the thrusting force to the south seems to have

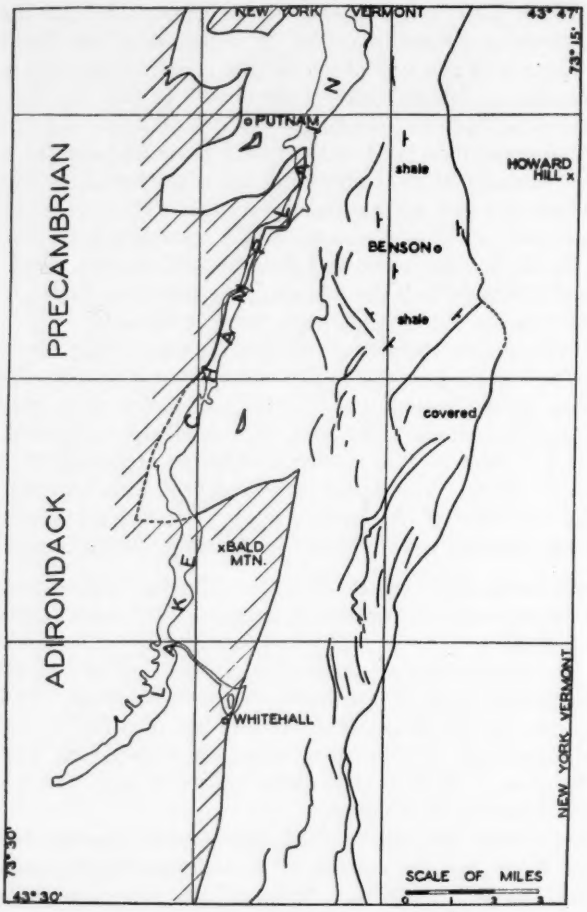

FIGURE 4.—*Map of thrust faults of the Whitehall quadrangle*

been general in the Taconic orogeny. Thus the Taconic thrust-plate reaches only a little farther north than this area, and it overlaps the eastern sequence rocks. These constitute another thrust-plate reaching to the latitude of Burlington and overriding the central sequence rocks,

where present. The central sequence rocks are another thrust-plate extending beyond the Canadian border, where they overlap the Philipsburg thrust-plate. All these sequences are thrust over the western sequence rocks. To the south in the Taconic quadrangle,[24] at the Massachusetts line, the slates in the Taconic thrust-plate are in turn overridden by one or two thrust-plates.

All the minor thrust faults of this area, with one exception, are small, the throw not exceeding 300 feet, unless great thicknesses of shale are involved. The exception is the fault already mentioned in connection with the easternmost longitudinal high-angle fault; it has a throw of 1500 feet. It is also the only thrust-fault that involves beds lower than Division D of the Calciferous. Apparently the higher, less resistant beds sliced along many small faults; the lower, more resistant dolomites broke only once. It is impossible to follow this fault into the northern third of the Whitehall quadrangle, because it is cut out by high-angle faults, but north of the northernmost transverse fault there is a fault with barely 500 feet throw. Traced north into the Ticonderoga quadrangle, however, the throw of this fault rapidly increases until it reaches 1700 feet just south of Shoreham. It probably represents therefore, if not the same fault, the resolution of the same forces.

These faults are cut in many places by all sets of the high-angle faults. Thus the faulting into blocks is later than the beginning of the Taconic orogeny, as has been suggested.[25] But if the blocks were overridden by the Taconic thrust, the high-angle faulting must have been over before the end of the orogeny, and must have occurred at a time when the area was under stress from the east.

SUMMARY

This area is important from a structural viewpoint because here the Adirondack high-angle fault structure and the Taconic thrust-fault structure approach closest together, and hence their age relations can best be worked out.

When the region was first subjected to orogenic forces, the western sequence rocks of the intermediate province were forced against the Adirondack massif. In the Whitehall quadrangle, where the belt is the narrowest, the rocks yielded by slicing; farther north where the belt is wider, as around Shoreham, there is considerable folding associated with the thrust faulting.

After this, the eastern part of the Adirondacks and much of the Champlain, upper Hudson, and Mohawk valleys fractured into great

[24] L. M. Prindle and E. B. Knopf: *Geology of the Taconic quadrangle*, Am. Jour. Sci., 5th ser., vol. 24 (1932) p. 257-302.

[25] G. R. Megathlin: Unpublished manuscript; to appear in N. Y. State Mus. Bull.

blocks. In this area, the orogenic forces, still acting from the east, pushed some of the east-west tiers of blocks past others, shattering them in the process.

Next, a great mass of slates was thrust from roots over 20 miles to the east near Rutland to within three miles of the Adirondacks.

Finally, the lessening forces, now from a direction more to the south, warped the region into a few irregularly distributed and broadly open folds trending north-northeast.

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GEOLOGY AND PETROLOGY OF EASTER ISLAND

BY MARK C. BANDY

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INTRODUCTION

Of the many volumes on the archeologic problems of Easter Island, only two [1] deal with the geology even in a fragmentary manner. Other

[1] Mrs. S. Routledge: *The mystery of Easter Island* (1920). London.
W. J. Thomson: *Te Pito Te Henua, or Easter Island*, Smithson. Instit. Rept. (1889) p. 447-552.

brief discussions are incomplete. Verlain [2] made the first attempt to describe the geology of the island, but his information was obtained from specimens deposited in France and his paper consists mostly of petrological descriptions. Skottsberg [3] described the general geology and

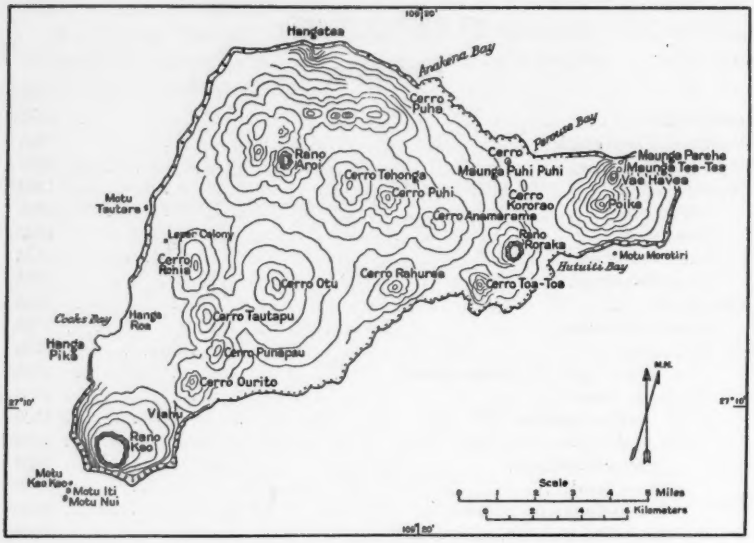

FIGURE 1.—*Map of Easter Island*

Modified from map of the Chilean Hydrographic Office. Contour interval, 50 meters.

physiography, after an examination in 1917. Chubb [4] visited the island in 1924 and has published the most comprehensive treatise on the geology and petrology. In addition, brief references are made to the petrology of the island by Rosenbusch,[5] Tilley,[6] Speight,[7] and Lacroix;[8] of these, the papers by Rosenbusch and Lacroix seem to contain errors which merit correction.

In the latter part of 1933, the writer devoted eight days to a study

[2] V. Verlain: *Les roches volcaniques de l'île de Pâques*, Soc. Géol. France, Bull., 3d ser., vol. 7 (1897) p. 410.

[3] Carl Skottsberg: *The natural history of Juan Fernandez and Easter Island* (1920) p. 4-8.

[4] L. J. Chubb: *Geology of Galapagos, Cocos, and Easter islands*, B. P. Bishop Mus., Bull. 110 (1933) p. 31-44.

[5] H. Rosenbusch and A. Ossan: *Elemente der Gesteinlehre*, 4th ed. (1923). Stuttgart.

[6] C. E. Tilley: *Density, refractivity and composition relations of some natural glasses*, Min. Mag., vol. 19 (1922) p. 276-294.

[7] R. Speight: *Note in:* J. M. Brown: *The riddle of the Pacific* (1924) p. 66-67. London.

[8] Alexander Lacroix: *La constitution lithologique des îles volcaniques de la Polynésie australe.* Acad. Sci. Paris, Mém., vol. 50 (1927) p. 1-82.

of the geology of the island and the collection of a large suite of rocks, which were studied petrographically at Harvard University. The map (Fig. 1) is modified from the map of the Chilean Hydrographic Office.

POSITION AND TOPOGRAPHY

Easter Island, or Rapa Nui, is situated in the Pacific Ocean, about 2300 miles west of Valparaiso, Chile. The position of the native settlement, Hanga Roa, on the west side of the island, is 27°10′ south latitude and 109°26′ west longitude. The island is on the southeast corner of the Albatross Plateau of the southeastern Pacific. It is about 15 miles long and 6 miles wide (maximum), and has an area of about 55 square miles. It is roughly triangular and rather rugged, with numerous small volcanic cones rising from the broad slopes of the larger cones. The cone, Rano Aroi, in the northwest corner, is the highest point (530 meters). In the southwest corner, Rano Kao rises to 316 meters, and on the eastern headland, Poike reaches a height of about 412 meters. These three volcanic piles dominate the topography.

The nearest landmass is the island of Sala y Gomez, composed of two barren volcanic peaks which rise a few feet above the waters of the Pacific about 210 miles east by north. Tahiti lies 2200 miles to the west, and the island of Juan Fernandez, off the coast of Chile, about 1800 miles to the east.

PHYSIOGRAPHIC HISTORY

The physiography of Easter Island is somewhat different from that of most of the volcanic islands in the South Pacific, in that there is practically no stream erosion. A single stream is normally fed by the overflow from the crater lake of Rano Aroi and flows southward only a few miles before it sinks into the porous mantle of volcanic ash. During heavy rains this stream evidently carries sufficient water to reach the ocean. At one time, practically its entire course was underground, but today the cover of the underground channel has caved near the headwaters, forming a deep gorge with many natural bridges. The other volcanic cones of the island have radial drainage. There is no visible overflow from either of the other two crater lakes.

Easter Island shows effects of disintegration rather than erosion. Under prevailing oxidizing conditions the mantle of ash becomes bright red. It is extremely porous, and even heavy rains immediately sink into the ground with a minimum of run-off, reappearing as springs at shallow depths along the coast. During his visit to the island the writer saw natives collecting this water for domestic use.

The island lies in the trade-wind region and has a fairly uniform climate throughout the year. Temperatures over 80° F. or under 55° F. are rare. Rain falls in all months, chiefly in showers or drizzles; it is variable but is probably between 45 and 50 inches a year, barely sufficient to support vegetation. The natives have destroyed almost all the trees; consequently, only coarse grass remains to hold the moisture in the ground, and even this has difficulty in attaining and keeping a foothold on the steep sides of some of the ash and volcanic cones and on the surface of the barren, rough flows. Even on the gentler slopes, there are broad areas of bare red ash.

Stream erosion has been slow. The marginal slopes are precipitous as they rise from the surface of the plateau, and, as wave erosion has been slow, there are few beaches and no good harbors. There is one small bay on the north coast—Anakena Bay. It is several hundred feet deep, about 200 feet wide, and has a small beach of shell sand, ten per cent of which is volcanic ash.

Wave erosion is actively undercutting the outer limits of the three main volcanoes, and has developed steep to precipitous cliffs at the three corners of the island. The most striking present-day example of this is on the south side of Rano Kao, where a vertical cliff over 1000 feet high has been developed. The rim of the crater has been lowered and, unless the island is uplifted, the impounded lake will eventually pour into the ocean. Cliffs on the other two corners are much lower and, on the whole, less spectacular. Wave erosion about the outer base of Poike has exposed the central vent of the volcano Maunga Parehe. If the cliff were less precipitous, this would be an excellent opportunity to study the core of a volcanic cone, as it is almost bisected to a depth of several hundred feet.

The tuffaceous cone of Rano Roraka now stands about a kilometer from the present shore line. It was formerly eroded by wave action similar to that active at the base of Rano Kao today. The erosion features are well preserved and shed light on a part of the geologic history of the island.

There were two major uplifts of the island, the more recent taking place in two stages. When the southeast side of Rano Roraka was being undercut by the waves, the ocean extended across the island and isolated Poike and the northeastern part of the island. At this same stage, Poike was undergoing wave erosion on its western side. This was a period of considerable duration and was apparently ended by an uplift of about 8 feet.

After a short interval the second major uplift, of two stages, began. Between the two major periods of uplift, the ocean extended across the

FIGURE 1. SOUTH COAST OF EASTER ISLAND
Looking southwest from the southwest slope of Rano Roraka. Pointed cone is Cerro Toa-Toa; the headland behind it is Rano Kao.

FIGURE 2. OUTER SLOPE OF THE CRATER OF RANO RORAKA

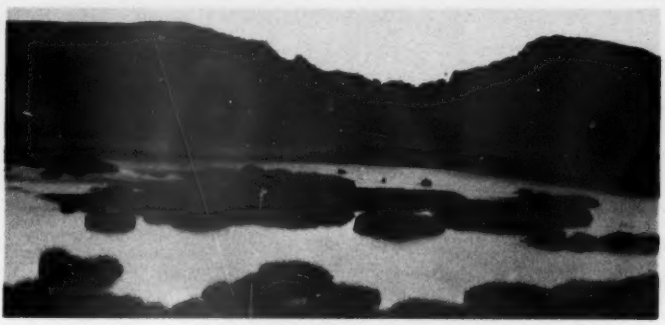

FIGURE 3. INNER CRATER OF RANO RORAKA
Showing effects of wave erosion on the rim and outer slope. Images set up on the inner crater can be seen. The wave terraces are not visible. All of the rock is the Rano Roraka tuff.

FIGURE 1. WAVE EROSION EFFECTS ON RANO KAO
Looking southwest across the crater and lake. The effect of present wave erosion in lowering the rim of the crater is well shown. The bedding of the lava flows is reflected in the profile of the rim. Beyond the rim, there is a sheer drop of 1000 feet to the ocean.

FIGURE 2. CENTRAL PLUG OF MAUNGA PAREHE
Central plug (the high point) and flows (beyond) on the north shore. The central plug of the volcano has been outlined by erosion.

island between Poike and Rano Roraka as a channel from 10 to 30 feet wide, the bottom of which was evidently lowered by wave-scour. The channel which can be easily traced today, at the foot of the gentle slope of Poike, is known to the natives of the island as the "Ditch of the Hanau Epe".

The first stage of the second uplift was about 8 feet. The ocean at that time completely encircled the base of Rano Kao, isolating it completely. At this time a small notch was cut in the base of Rano Roraka. The final stage of the uplift, about 12 feet, caused a retreat of the ocean to about its present position.

An uplift of 30 feet today would probably expose an area to the south and west of Rano Kao, similar in every respect to the area to the northeast of the volcano and to the north of Rano Roraka.

GENERAL GEOLOGY

INTRODUCTION

Easter Island is entirely volcanic and was built up in three stages, the parts formed in succession being: (a) the extreme east; (b) the central region, including the two volcanoes Rano Aroi and Rano Roraka; and (c) the southwestern region, including Rano Kao and a number of isolated cones. The eruptions were all basaltic with the exception of local eruptions during the last period in the western part of the island.

POIKE STAGE

The first cone to emerge from the Pacific was Poike. Here the soil is deepest although the bed rock is as dense as any of the basaltic flows. The soil has completely covered all the flows that issued from this cone. There is no trace of a crater, and yet there is no reason to believe that one did not develop similar to those of the other large volcanoes on the island. At one time the cone extended farther, especially to the north, east, and south, as a more or less symmetrical cone. Wave erosion has greatly reduced its former symmetry.

The rocks of this eastern area are distinctive in their general character, although the mineral composition is only slightly different from that of the rocks of most of the island. The disintegration of the lava, tuff, or ash has been so complete that sound rock is scarce.

The cliff sections show ash beds and tuffs commonly interbedded with the flows. The flows are seldom over 10 feet thick and are more or less uniform in composition and type. No evidence could be found to indicate that any of these exposed rocks were of subaqueous origin.

RANO AROI STAGE

The next cone to take form was the main large and complex area that is here called Rano Aroi, after the main volcanic mass. At present it

extends south to Cerro Rohia, Cerro Punapau, and Cerro Rahurea; east to Hutuiti Bay and Perouse Bay; and north and west to the ocean. Wave erosion has removed an unknown greater extent to the north and west. Mineralogically the Rano Aroi rocks are the same as those of Poike. Near the close of the period of formation of the main Rano Aroi cone, many small satellites were formed on its slopes, and these poured out a comparatively large amount of lava.

Two rows of volcanoes appear to occupy radial lines of weakness on Rano Aroi, and there may be others. One row includes the cones of Cerro Tehonga, Cerro Puhi, and Cerro Anamarama; the other row is made up of a number of small unnamed cones north of Rano Aroi. Rano Aroi contains a crater lake that is not large now and probably never was as large as the other crater lakes on the island. As mentioned before, the overflow of this lake has cut a small gorge and underground channel in the ash and tuff on the south side of the crater. Where the underground channel is still preserved, the openings are large enough to admit a man on horseback.

In the exposed sections on the north side of Rano Aroi, some flows are over 50 feet thick, and some are probably even thicker. They are mainly coarse-grained porphyries.

It would appear that toward the end of the Rano Aroi stage, Rano Roraka began to form and build up a cone out of palagonite tuff and ash. Some lava was poured out at a very early stage in this development. This cone was previously probably somewhat higher. Wave erosion has cut a notch in the southeast portion of the rim, and disintegration has lowered and rounded the northern rim. After the activity that marked the formation of Rano Roraka, the ocean invaded the island and surrounded the base of this cone, as by a moat. The strip of land that was flooded is comparatively free from ash and tuff today, and abounds with rough exposures of flows that issued from Rano Aroi or some of the satellites. The time since the ocean withdrew from the land has not been sufficient to disintegrate this lava surface. With the exception of the area surrounding Rano Kao, Cerro Ourito, and Cerro Otu, this is probably the roughest area on the island.

RANO KAO STAGE

The next mass to take form produced Rano Kao and other smaller cones in its vicinity. Rano Kao was built up to about 1050 feet, and the rim has since been little lowered by erosion. One cannot tell whether Cerro Ourito, Cerro Otu, and other cones in this vicinity are earlier than, later than, or contemporaneous with the eruption from Rano Kao. The youngest extrusive, the Cerro Ourito obsidian, was poured out in

greatest abundance, and in about equal amounts, from Cerro Ourito, Cerro Otu, and Rano Kao.

An early volcanic product of this stage is the Punapau pumice. This pumice forms a low cone, Cerro Punapau, 100 feet high and 300 feet wide at the base. It is separately designated and described because of the important place it occupies in the archeology of Easter Island. Cerro Punapau, immediately west, and at the foot, of Cerro Otu, probably marks one of the first points of activity of the Rano Kao period. The lack of wave erosion is evidence that it did not belong to the Rano Aroi stage. Its life was short, and its activity ceased long before that of Cerro Otu.

The marked feature of the Rano Kao stage is the constant decrease in the amount of volatiles in rocks of succeeding eruptions. The early activity, as illustrated by the Punapau pumice cone, produced a highly porous rock. This pumice can be duplicated in other parts of the island. Succeeding outbursts show increasingly smaller amounts of volatiles, and the flows become proportionately less vesicular. On Cerro Otu the rocks range from typical early, coarsely vesicular basalts to the black dense microvesicular later lavas characteristic of this cone. Above the lavas is a small amount of glassy obsidian showing no vesicles. The flows on Cerro Ourito grade up from normal vesicular types to a long-continued series of flows of dense obsidian, some of which, however, is viscous and ropy and contains elongated vesicles. The flows from Rano Kao are similar in every respect to those of the two cones already described.

During the Rano Kao stage, three cones were formed on the extreme northeast side of the island. These form the Maunga Parehe area. According to clear field evidence, these two groups of cones at the two extremes of the island were formed contemporaneously and at the latest stage of vulcanism. That conclusion is amply borne out by the petrology of the two areas. Maunga Parehe, Maunga Tea-Tea, and Vae Havea have histories slightly different from that of the Rano Kao group of volcanoes, but the products of both groups are practically the same. No obsidian was found in the Maunga Parehe area, but this is the only significant difference. None of the three cones contains a crater. A good spring issues from a deep cleft in Maunga Tea-Tea, and another supplies the water to fill the gargantuan mouth of a markedly negroid head, carved out of the steep southern slope of Vae Havea.

The writer believes that both Rano Kao and Maunga Parehe tapped essentially the same magma source more or less simultaneously. The flows from the vents of the Maunga Parehe area began later than from Rano Kao and did not continue as long.

PETROGRAPHY

GENERAL STATEMENT

The following rock types are believed to be distinctive enough to merit description. Average compositions and textures are given. In many cases, groups of rocks that vary slightly mineralogically and texturally are included in a single group. As a rule, the different groups are surprisingly uniform in their various physical and microscopical character. The rock type of many of the cut blocks of stone used in the ahus, the local name for the old burial cairns, and house foundations can be recognized by the distinctive physical and megascopical characters. As nearly as possible, these rock types are given in the order of their age, from the oldest to the youngest.

Due to the vesicular character of the rocks of Easter Island, specific gravity determinations would have little meaning.

POIKE BASALT

The Poike basalt is dark gray, with medium to coarse vesicles ranging in size up to 5.0 mm. The texture is aphanitic with scattered phenocrysts of plagioclase and olivine.

The plagioclase phenocrysts range from Ab_{18} to Ab_{25}, and up to 5.0 mm. in longest dimension. They are strongly zoned and show Carlsbad and albite twinning. Olivine rarely occurs as phenocrysts and never as large grains. Magnetite is likewise rare as large grains. The groundmass is made up of plagioclase laths (Ab_{50} to Ab_{45}) with interstitial pigeonite, olivine, magnetite, and a small amount of glass. There are small grains of brown augite with the following optical properties: $n_\beta = 1.700 \pm .005$, $2V = 60°$, $Z \wedge c = 45°$. Apatite varies in quantity though never abundant. Iddingsite may replace much of the olivine in some of the flows, especially those on the north shore. Magnetite is abundant; skeleton crystals occur in all the flows and associated with pyroxene are best developed in the glass around vesicles. The feldspar from these areas shows extreme replacement. Halloysite is a common alteration product of the plagioclase phenocrysts. It crosses the plagioclase in replacement veinlets that in many places widen out in the core of the larger phenocrysts, where alteration is extensive.

The characteristic features of this type are the aphanitic groundmass with few phenocrysts and the relatively large and more or less spherical vesicles, free from any minerals. The dark color, combined with the characteristic vesicles, is a relatively distinctive field marker.

RANO AROI PORPHYRITIC OLIVINE BASALT

This rock type makes up about 30 per cent of the surface of the island. Its most typical development is about the vent, around the crater lake,

and on the shore to the north of Rano Aroi where it forms high, steep cliffs. It is medium to dark gray and medium to coarsely vesicular. It is conspicuously porphyritic with large phenocrysts of glassy plagioclase, olivine, and augite. As a rule, the groundmass is megascopic. The vesicles are lined with hematite in bright glistening crystals, and possibly with limonite.

The plagioclase phenocrysts make up about 20 per cent of the rock and range up to 10 mm. in length. The average composition is from Ab_{20} to Ab_{35}. The phenocrysts of olivine range in size up to 3 mm. and vary considerably in optical properties: $n_\beta = 1.715$ to 1.742, $2V = 90°$ to 75°. There is a general relationship between the composition and the size of the olivine crystals. The larger crystals are richer in fayalite molecule. The composition of the olivine, as determined from the optical properties, varies from Fo_{57} to Fo_{68} and, although iron-rich, is far from fayalite. The analysis of this rock (Table 1, No. 7) shows a normative olivine of composition Fo_{49}. The olivine grains are clear, rounded, usually in clusters, and distinctly brown megascopically. The augite phenocrysts are free from zoning, and twinning is not common. They are brown, and no pleochroism was observed. Optically, the augite shows, $n_\beta = 1.703$, $2V = 60°$. The extinction angle varies somewhat in different specimens, but the indices are constant.

The groundmass is microgranitoid. The plagioclase grades from phenocrysts into the groundmass with no marked break, either in size or in composition, and, in the groundmass, ranges from Ab_{35} to Ab_{50}. Olivine of the approximate composition Fo_{65} occurs in minor amounts and is locally altered slightly to iddingsite. Pyroxene of both the diopside-hedenbergite and the pigeonite types occurs in different specimens and even in the same specimen. The former type is present only in small phenocrysts. The pigeonite has $2V = 20°$, and is pleochroic, X—yellowish, Z—very pale green. Apatite is present in minor amounts. Both granular and arborescent or skeleton magnetite are present. Some of the plagioclase phenocrysts contain magnetite, but this is not a characteristic feature of the rock. The plagioclase phenocrysts, especially the cores, have been strongly or entirely altered to halloysite and some isotropic mineral, not identified.

This rock type is essentially a coarse-grained equivalent of the Poike basalt. The characteristic features are the coarse texture and the presence of large phenocrysts of augite. The analysis shows it to be a typical olivine basalt. Due to the silica deficiency, olivine is present in the normative minerals in great excess over hypersthene. Compared with the average analysis for basalt, as calculated by Daly,[9] the Easter

[9] R. A. Daly: *Igneous rocks and the depths of the earth* (1933) p. 17. New York.

Island rock is seen to have a high percentage of titanium oxide and soda, and a low magnesia content. Many analyses of Hawaiian rocks that have been published [10] are very similar to this type. Due to its vesicular and coarse texture, the rock was easy to work by primitive method, and most of the foundation stones of the "canoe houses" of the old inhabitants were made from this rock, as well as many of the cut and fitted blocks used about the ahus.

TONGARIKI BASALT

The Tongariki basalt group includes a large number of flows from many small cones about Rano Aroi and between Rano Aroi and Poike, varying in mineral composition but essentially similar. The color varies from light to medium gray. All are vesicular, but the vesicles are smaller than in the two types just described, and more irregular, and hematite and ilmenite may occur as linings. The analyzed rock from this group (Table 1, No. 6) was somewhat porphyritic, with a few phenocrysts of plagioclase and an occasional olivine grain in a fine groundmass.

Plagioclase, with phenocrysts up to 5.0 mm., is the most important phenocryst mineral. Its composition, surprisingly uniform, varies from Ab_{80} to Ab_{55}. The distinctive feature of the olivine is its occurrence in elongated, wedge-shaped crystals or grains. These elongated crystals are never complete units but are crossed by veinlets which break the crystals into several parts. Quensel [11] describes and pictures a similar form of olivine from Isla Mas a Tierra. The composition of these phenocrysts does not vary from that of the small grains present in this group of rocks and in others of the island. Though not confined to one rock group, this type of olivine is more commonly developed here than in other rocks. The optical properties are, $n_\beta = 1.707$, $(-)2V = 85°$. These correspond to an approximate composition of Fo_{75}. The normative olivine as determined from the analysis of the rock has a composition of Fo_{58}. It may be noted that, here again, the normative mineral contains a much higher percentage of the fayalite molecule than the olivine actually in the rock.

The groundmass is very fine grained, as a rule, with variable amounts of glass. Laths of plagioclase predominate; they vary in composition from Ab_{75} to Ab_{50}. Generally the rocks show either a diabasic or a trachytic texture, although in some of the flows there is so much glass that the plagioclase is limited to very small grains. The interstices

[10] H. S. Washington: *Petrology of the Hawaiian Islands*, Am. Jour. Sci., 5th ser., vol. 6 (1923) p. 338-365.

[11] P. D. Quensel: *Die Geologie der Juan Fernandez Inseln*, Geol. Instit. Upsala, Bull. 11 (1912) p. 263-264.

of the net of plagioclase crystals are filled with pigeonite (2V about 25°), olivine, apatite, magnetite, iddingsite, and glass. The olivine in the groundmass of this group is the most magnesia-rich of any examined. Its index varies, $n_\beta = 1.65$ to 1.70, which shows a wide variation in composition. The average corresponds to a composition of about Fo_{90}. The stated variations in composition represent the group and not individual specimens which show only small variations. Generally, however, olivine in the groundmass contains a lower ratio of fayalite than that in the phenocrysts. In some flows of this group, the amount of olivine approximates that of pyroxene, but as a rule the pyroxene predominates. A few grains of brown augite were observed. These are negative ($2V = 45°$ to 50°) and in some of the flows show a tendency to develop hour-glass structure. Magnetite occurs both in grains and in skeleton crystals. In the two types previously described, skeleton crystals are always present, but in this group many flows do not carry a single skeleton crystal; actually, only about 35 per cent of the flows studied bear skeleton crystals. The granular magnetite is characteristically fine-grained as contrasted with both earlier and later rock groups. With rare exceptions, all the magnetite is finer than 0.05 mm.

Apatite is present in about half of the rocks but is rarely abundant. Brown tachylitic glass with an average index of 1.545 is almost always present. Iddingsite is present in most of the flows, although it is seldom conspicuous. It commonly replaces olivine grains at the core, leaving a narrow unreplaced rim. Halloysite is rare.

The flows from Rano Roraka are taken as typical of this class, and the material analyzed came from one of them, between Rano Roraka and the Ahu Tongariki, from which the group derives its name. It shows phenocrysts of plagioclase (Ab_{30}) and elongated olivine grains (Fo_{75}). Iddingsite is present in small amounts. Magnetite occurs in grains, and pigeonite is strongly dominant over olivine in the groundmass. The rock analysis shows a silica content about the same as that of the Rano Aroi olivine basalt. Nevertheless, the rock is distinctly more salic. In calculating the normative minerals, the analysis was found to be delicately balanced between a type with normative olivine and one with normative quartz. The norm shows a much higher percentage of pyroxene than the rock actually contains and, of course, fails to show the 10 per cent or more of olivine in the rock; it reflects the increase in sodic and potassic feldspar in the rock.

RANO RORAKA TACHYLITE TUFF

A light, porous tuff forms the crater of Rano Roraka, and all the great images on Easter Island are carved from it. It contains many inclusions ranging in size up to one meter, some of which have been partially

TABLE 1.—*Analyses of Easter Island rocks* *

	a	b	c	1	1a	1b	2	d	e	f	3	g
SiO_2	74.22	72.86	72.74	72.51	74.24	72.32	71.92	68.50	66.58	60.92	59.75	54.88
TiO_2	0.09	0.19	0.28	0.21	0.18	0.21	0.28	0.22	0.42	1.88	0.70	2.28
Al_2O_3	12.25	13.48	9.14	13.35	13.44	13.34	13.59	14.91	14.55	13.47	15.42	13.54
Fe_2O_3	0.40	2.34	2.79	1.35	0.08	1.49	0.51	3.72	2.35	2.99	3.47	2.74
FeO	2.52	0.81	0.80	1.99	2.16	1.97	2.71	0.77	2.72	5.57	7.03	10.08
MnO	0.06	0.03	0.05	.....	.....	.....	0.11	0.05	0.06	0.14	0.38	0.12
MgO	tr.	0.06	0.10	0.16	0.16	0.16	0.07	tr.	tr.	1.65	0.93	2.37
CaO	0.98	0.36	0.94	0.47	0.54	0.46	1.08	0.54	0.96	4.98	4.50	6.48
Na_2O	5.21	5.44	4.42	4.81	5.81	4.70	5.93	4.63	5.93	4.61	4.75	4.14
K_2O	3.82	3.46	2.82	4.48	2.79	4.66	3.37	3.07	3.92	1.73	1.74	1.27
H_2O-	0.16	0.60	2.72	.....	.....	.....	0.11	1.37	0.75	0.83	0.50	0.74
H_2O+	0.67	0.86	2.33	0.47	0.34	0.48	0.18	2.69	1.69	1.01	0.35	0.45
CO_2	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
P_2O_5	tr.	0	1.11	.....	.....	.....	.....	tr.	tr.	0.35	0.34	0.74
S	.....	.....	.....	0.05	.....	0.06	.....	.....	.....	.....	tr.	.....
BaO	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
	100.43†	100.49	100.24	99.85	99.74	99.85	99.86	100.47	99.93	100.13	99.86	99.83
Norms												
q	27.06	27.12	37.62	25.44	26.75	25.14	21.42	28.74	15.06	15.78	12.18	7.44
or	22.24	20.57	16.68	26.69	16.68	27.80	20.02	18.35	22.80	10.01	10.01	7.78
ab	41.92	46.11	30.92	40.35	49.25	39.82	49.78	38.77	49.78	38.77	40 35	35.11
an	.....	1.67	.....	1.67	1.95	1.11	0.56	2.78	1.67	11.12	15.57	14.18
di	4.47	.....	.....	0.49	0.75	0.96	4.43	.....	2.48	9.93	3.84	11.60
hy	2.24	.....	0.33	2.38	3.44	1.88	2.08	.....	1.32	3.91	9.82	12.52
ac	1.39	.....	5.54	.....	.....	.....	.....	.....	.....	.....	.....	.....
ol	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
mag	.....	1.86	1.39	1.86	0.23	2.09	0.70	2.09	3.25	4.41	5.10	3.94
il	0.15	0.46	0.61	0.46	0.46	0.46	0.61	0.46	0.76	3.65	1.37	4.41
ap	.....	.....	1.68	.....	.....	.....	.....	.....	.....	0.67	0.84	1.68
hm	.....	0.96	.....	.....	.....	.....	.....	2.24	.....	.....	.....	.....
P_2O_5	.....	.....	0.40	.....	.....	.....	.....	.....	.....	.....	.....	.....
Na_2SiO_3	0.12	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
C	.....	0.40	.....	.....	.....	.....	.....	2.96	.....	.....	.....	.....

* Analyses lettered "a" to "k" [taken from Alexander Lacroix: *Composition chimique des laves de l'île de Pâques*, Acad. Sci. Paris, C. R., vol. 202 (1936) p. 601–605] were added to the table by the editor, as Lacroix' papers were not available to the writer.

† Cl 0.05.

a. Obsidienne hyperalcaline rubanée, Rano Kao, analyzed by Raoult.
b. Rhyolitoïde hyperalcaline rubanée, Motui Nui, analyzed by Raoult.
c. Rhyolitoïde hyperalcaline rubanée, Motui Nui, analyzed by Raoult.
1. Cerro Ourito obsidian, Rano Kao, No. A–IV–c–2–1, analyzed by F. A. Gonyer.
1a. Spherulites in the Cerro Ourito obsidian, Rano Kao, No. A–IV–c–2–1, analyzed by F. A. Gonyer.
1b. Glass without spherulites, Cerro Ourito obsidian, Rano Kao, No. A–IV–c–2–1, analyzed by F. A. Gonyer.
2. Rhyolitic obsidian (Cerro Ourito obsidian), Mt. O(u)rito, analyzed by Raoult.
d. Rhyolitoïde à aspect crayeux, Mont Orito, analyzed by Raoult.
e. Rhyolitoïde hyperalcaline rubanée, Poike (statue), analyzed by Raoult.
f. Dacitoïde oligoclasique, Rano Kao, analyzed by Raoult.
3. Andesite, Mataveri, St. George collection No. 862, analyzed by Herdsman.
g. Basalte aphyrique α (block dans brèche), Rano Roraku, analyzed by Raoult.

	4	5	h	6	7	i	j	k	8	8a	l	9
SiO_2	50.02	49.67	48.34	48.14	48.02	46.66	46.06	45.68	45.52	51.81	42.92	40.12
TiO_2	2.42	.....	3.78	3.96	3.34	4.06	4.50	4.06	2.40	2.73	4.68	.....
Al_2O_3	15.28	14.46	13.06	15.00	17.08	14.28	14.70	14.13	14.32	16.30	15.70	13.27
Fe_2O_3	1.82	18.52	1.96	4.38	2.43	3.97	4.04	10.20	6.92	7.87	7.84	10.65
FeO	10.86		10.08	10.15	9.87	8.78	7.96	2.66	5.14	5.85	7.56	
MnO	0.20	.....	0.19	0.06	0.12	0.12	0.15	0.10	0.20	0.23	0.17	.....
MgO	3.31	3.74	4.74	4.63	4.58	5.35	4.86	5.50	2.98	3.40	3.66	3.32
CaO	8.08	7.23	10.40	8.56	9.40	7.54	10.82	10.10	5.88	6.69	8.22	9.47
Na_2O	4.42	2.92	3.61	3.34	3.22	2.57	2.82	3.64	2.89	3.29	3.47	2.06
K_2O	1.37	1.64	0.90	1.00	0.78	1.00	0.79	0.77	1.34	1.52	0.67	0.97
H_2O-	0.33	1.17	1.02	.....	.....	2.14	0.82	0.47	7.22	.....	1.73	20.43
H_2O+	0.81		1.37	0.84	0.86	3.35	1.81	2.68	4.81	.....	2.64	
CO_2	0.70	.....	.....	.....	.....	.....	.....	.....	0.30	.....	.....	.....
P_2O_5	0.40	.....	0.47	0.08	0.04	0.47	0.50	0.45	0.27	0.31	1.03	.....
S	.....	.....	.....	0.05	0.02	.....	.....	.....	.....	.....	.....	.....
BaO	.....	.....	.....	0.02	0.03	.....	.....	.....	.....	.....	.....	.....
	100.02	99.35	99.92	100.21	99.79	100.29	100.13	100.44	100.19	100.00	100.29	100.29
Norms												
q	.....	.....	.....	0.43	.....	4.20	1.44	.....	8.82	10.09	1.86	.....
or	8.34	.....	5.56	6.12	5.00	6.12	5.00	4.45	7.76	8.90	3.89	.....
ab	37.20	.....	30.39	28.30	27.25	22.01	23.58	30.92	24.63	28.17	29.34	.....
an	17.79	.....	16.68	22.80	29.75	24.19	25.02	19.74	21.96	25.11	25.30	.....
di	12.90	.....	25.91	15.46	13.98	7.94	20.02	18.18	2.81	3.21	7.13	.....
hy	2.26	.....	1.92	12.35	6.80	15.62	6.22	1.90	6.23	7.12	5.80	.....
ac	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....	.....
ol	10.78	.....	5.80	6.34	.....	.....	.....	2.66	.....	.....	.....	.....
mag	2.55	.....	3.02	6.50	3.48	5.80	5.80	.....	9.98	11.41	11.37	.....
il	4.56	.....	7.30	7.60	6.23	7.75	9.12	5.78	4.56	5.22	8.97	.....
ap	1.01	.....	1.34	0.34	.....	1.34	1.34	1.01	0.67	0.77	2.35	.....
hm	.....	.....	.....	.....	.....	.....	.....	10.24	.....	.....	.....	.....

4. Andesitic basalt, fragment from andesitic tuff, probably Rano Roraka tuff (from Mataveri according to Lacroix), analyzed by Raoult.
5. Sideromelane, locality not recorded, reported in Rosenbusch.
h. Basalt aphyrique β (en plein), Tonga Riki, analyzed by Raoult.
6. Tongariki basalt, No. D–II–d–4, Rano Roraka, analyzed by F. A. Gonyer.
7. Rano Aroi porphyritic olivine basalt, No. B–I–e–1, on the coast north of Rano Aroi, analyzed by F. A. Gonyer.
i. Dolérite, Rano Kao, analyzed by Raoult.
j. Basalte porphyrique à labrador, Baie de Lapérouse, analyzed by Raoult.
k. Basalte β (scorie), Anakana, analyzed by Raoult.
8. Andesitic tuff, probably from Rano Roraka [Mataveri according to Lacroix] (Rano Roraka tachylyte tuff), analyzed by Raoult.
8a. Same as 8 but calculated on a water-free basis.
l. Basalte aphyrique α, Baie de Lapérouse, analyzed by Raoult.
9. Palagonite, locality not given, reported in Rosenbusch.

melted and resorbed; others have sharp contacts with a very light bonding. Many times those who carved the images were compelled to abandon a partially completed image when one of these large and hard xenoliths was encountered. The inclusions were turned to a good use, however, in being made into stone hammers and chisels, for which they are admirably suited on account of their density and hardness.

The tuff weathers to a rough surface, the harder bands standing up in relief above the purer tachylite pumice. The rock probably disintegrates very slowly, as modelled features of many of the images are still sharp and clean today. The tachylite glass has an index of $1.566 \pm .002$, corresponding to a silica content of about 52 per cent[12]; an analysis by Raoult shows the rock to contain approximately this amount of silica (Table 1, No. 8a). The rock is greenish yellow on a broken surface but weathers to a yellowish gray. The tachylite is so porous that much of it has a specific gravity under 1.7. Some of the small microvesicles are filled with a fibrous mineral regarded as a zeolite, but not accurately determined. Phenocrysts of pyroxene, olivine, and plagioclase are not common. Some of the smaller feldspar laths have the approximate composition of Ab_{30}; they grade up to the most sodic type observed, Ab_{47}. The pyroxene is augite. Olivine occurs in the peculiar elongated grains previously described and also in coarse grains or phenocrysts. Magnetite, in fine grains, and iddingsite are also found.

Weathering accentuates the stratification in the tuff. On the whole, the banding is due to changes in grain size, and the beds are thin, usually less than 5.0 cm. This stratification is often cited as evidence of sub-aqueous deposition of the tuff. Speight[13] details the lack of evidence for its submarine eruption.

ROCKS OF THE RANO KAO STAGE

General Statement.—The last, Rano Kao, stage of volcanism is represented by a number of rock types. This outpouring was confined to the southwestern part of the island, about Rano Kao. Five different rock types in this area have been studied, Viahu basalt, Cerro Otu olivine andesite, Rano Kao andesite, Cerro Punapau pumice, and Cerro Ourito obsidian. Another rock group, Maunga Parehe olivine dacite tuff, has been included in the period, on the basis of field evidence and of chemical and mineralogical similarity, though it may possibly have been formed at the end of the Rano Aroi period.

The rocks vary from normal basalts through andesites to obsidians. This period may have been no longer than that of the Rano Aroi period,

[12] W. O. George: *The relation of the physical properties of natural glasses to their chemical composition*, Jour. Geol., vol. 32 (1924) p. 353-372.

[13] R. Speight *in* J. M. Brown: *The riddle of the Pacific* (1924) p. 66-67. London.

but differentiation was more active. In some places but at different times the same vent poured out basalt, andesite, and obsidian. At Rano Kao, basalt was the earliest rock extruded and obsidian the last.

Viahu Basalt.—It was impossible to obtain a specimen of an early flow from Rano Kao, in place. The beach about the volcano was made up of basaltic boulders, undoubtedly broken from early Rano Kao flows. A quarry in a later flow yielded the dressed stones in the walls of the Ahu Viahu. This flow, although not one of the earliest of the period, was the oldest studied. The rock is medium gray and aphanitic. The vesicles are small, irregular, and commonly lined with what is believed to be a form of iron oxide. The rock weathers deeply to a yellowish color. Phenocrysts are not abundant. They include scattered plagioclase crystals (Ab_{40}), and rare, small (0.5 mm.) augites with some tendency to show the hour-glass structure. Olivine occurs in elongated grains, as in the Tongariki basalt, having the general composition of Fo_{70} and grading down to small grains in the groundmass. As the grains become smaller, they become equidimensional. In the groundmass, plagioclase of the composition Ab_{60} is the predominant mineral. Associated with it are pigeonite and olivine. Magnetite is abundant in minute grains throughout the groundmass. The rock is surprisingly free from iddingsite.

Cerro Otu Olivine Andesite.—One of the late products of volcanism of Cerro Otu, formed in the Rano Kao period, is the Cerro Otu olivine andesite. It is very dense, black, and microvesicular, and gives a clear tone when struck with a hammer. Due to its hardness and texture, it was a favorite material for stone chisels. Those sectioned for this study were made of either this basalt or the olivine-rich basalt xenoliths in the Rano Roraka tuff. Most of this type occurs as ejecta on the side of the volcanic cone from which the name is derived. A few phenocrysts of plagioclase, Ab_{55}, occur in an aphanitic groundmass. This is notably a relatively sodic type of plagioclase and also a late rock type, closely related in time and place of eruption to the Cerro Ourito obsidian. The groundmass plagioclase may occur as scattered laths in a glassy magnetite-charged base, or may be more abundant, even to the development of a microdiabasic texture. The composition is essentially the same as that of the phenocrysts. Pyroxene ranges from normal augite to diopside-rich pigeonite. Hypersthene is possibly present. Olivine, commonly altered to iddingsite, is an accessory or subordinate mineral. Magnetite is locally very abundant as dust through the groundmass, increasing with the amount of glass. Apatite, nowhere conspicuous, is present only in the more crystalline phases. Glass may constitute from 15 to over 65 per cent of the rock volume. An unsatisfactory deter-

mination of the index of the glass gave a value of $n = 1.51$—below the value of 1.566 obtained for the Rano Roraka tachylite but above the 1.496 value obtained for the Cerro Ourito obsidian. In addition to magnetite dust, the glass is usually densely charged with minute grains of pyroxene, and then tends to be opaque, even in very thin sections.

Rano Kao Andesite.—The Rano Kao andesite makes the rim and the upper portion of the Rano Kao cone. Specimens were taken from flows, exposed by slumping, to a depth of 150 feet within the cone. The structural features of the rock are unique. Rounded bodies, commonly over 15 feet in diameter and with an onion-like structure, much like boulders of disintegration in granite, result from weathering. Because of the platy fracture, the rock can be broken into large flat pieces, 3 to 8 centimeters thick. All the houses of the old cult of "Bird Men" of Orongo, situated on the edge of the crater, are built of this rock.

There are no striking differences through a vertical range of over 150 feet. This was one of the most fluid lavas on the island, yielding ash-free slopes that are smooth and regular in contrast with those of other cones. The texture likewise reflects the liquidity of the lava. The vesicles are irregular and elongate, parallel to the direction of flow. Like the Viahu basalt, the weathered rock is yellowish, especially around the vesicles. These contain rare small (0.5 mm.) balls of cristobalite ($n_0 = 1.488$). The lower flows exposed in the crater show no cristobalite. Inclusions of older rocks in the flows are numerous.

Phenocrysts are common. Plagioclase varies slightly in composition (Ab_{50} to Ab_{55}), but some phenocrysts have cores as calcic as Ab_{35}. Pyroxene is rare; it forms small phenocrysts that give an extinction angle of $Z \wedge c = 45°$. Olivine is absent. The groundmass consists of a mat of plagioclase laths (Ab_{75}), with interstitial fine-grained masses of a pyroxene, glass ($n = 1.50$ or less), and magnetite. The pyroxene is apparently diopside and not pigeonite as in the older rocks ($2V = 55°$, $Z \wedge c = 40°$ to $45°$). Its grains are too minute to give accurate optical characters. Magnetite is present as fine dust or small grains through the groundmass.

Maunga Parehe Olivine Dacite Tuff.—This tuff is one of the most extraordinary rocks on the island. It forms the three peaks on the north slope of Poike, Maunga Parehe, Maunga Tea-Tea, and Vae Havea. Despite its apparent softness, it must weather slowly, for carvings made from it at least two hundred years ago are still sharp and clean. One wonders why this rock was not used in making the images.

The two most characteristic features are the abundance of cristobalite and the presence of olivine phenocrysts, now entirely altered to iddingsite. The tuff is very fine grained, compact, purplish gray, soft, and

FIGURE 1. IMAGE SHOWING BEDDING ACCENTUATED BY DIFFERENTIAL WEATHERING

FIGURE 2. IMAGE SHOWING AN INCLUSION OF TANKARIKI BASALT
Coarse Lapillae are evident.

FIGURE 3. IMAGE SHOWING DETAIL OF NASAL ORNAMENTATION PRESERVED FOR CENTURIES

RANO RORAKA TACHYLITE TUFF

FIGURE 1. CERRO PUNAPAU CRATER AND QUARRY
Taken from west side of Cerro Otu, looking west. Unfinished "hats" are seen in the quarry.

FIGURE 2. FINISHED "HAT"
Had been placed on an image of the Ahu Tongariki. The extreme porosity of the pumice is evident.

rather pulverulent, but is not markedly vesicular. Sharply euhedral phenocrysts of plagioclase are present, up to 1.5 mm. in size. Hand specimens also show abundant balls of cristobalite ($n_o = 1.488$) lining or filling vesicles. The phenocrysts of plagioclase have an average composition of Ab_{90}. They show fine albite and Carlsbad twinning and, as a rule, are sharply euhedral in form. Small phenocrysts of quartz, strongly resorbed and replaced, are present in the tuff. Olivine phenocrysts, now altered to iddingsite, are common. There is not enough unaltered olivine to enable optical determination of its composition, although it is believed to have been a fayalite-rich variety. There are scattered phenocrysts of an unusual triclinic feldspar with the index range of anorthoclase. These crystals are not clear units but appear only in outline in the groundmass, upon rotation of the stage with crossed nicols. They are identical in form with the phenocrysts found in some of the Cerro Ourito obsidian, and the index is assumed to be the same in this rock as in the obsidian. The groundmass consists of quartz, albite, and cristobalite filling and lining vesicles and extending into the rock. An alteration product scattered through parts of the rock is tentatively identified as a member of the hydrotalcite group.

Cerro Punapau Pumice.—Cerro Punapau is a small cone of pumice on the western side of Cerro Otu. All the hats or "maroke" of the images on the ahus came from this place. Partly or wholly completed hats are to be seen within and around the cone. The hat quarry proper was inside the cone. The pumice is coarsely vesicular, vesicles ranging up to over 15 cm. in diameter. The color varies from a bright red, on weathered surfaces, to almost black on fresh fractures. Much of it is ropy and scoriaceous. Phenocrysts are not evident megascopically.

Thin sections of the rock show it to be highly vesicular. The vesicles have extremely thin walls of black, more or less opaque glass. There are a few crystals of plagioclase of the composition Ab_{45}. The groundmass laths of plagioclase have the composition of Ab_{50}. A few grains of olivine, commonly altered to iddingsite, were seen, but no accurate optical data on them could be determined. Pyroxene, both augite and pigeonite, occur in the pumice. The augite variety tends to show hourglass structure. The larger grains of pyroxene are more abundant than olivine.

The mass of the pumice is a basaltic glass, heavily charged with iron oxide and microlites. Its apparent index is 1.577, but the glass is so opaque and there is so much iron oxide present that the determination is uncertain. Such an index would indicate a silica content of about 51 per cent. This rock type is only of interest from an archeological

point of view. However, much of the red surface "soil" of the island is of this same general composition.

Cerro Ourito Obsidian.—The last volcanic product formed on the island was the Cerro Ourito obsidian. This flow poured from the vents of Cerro Ourito, Rano Kao, and Cerro Otu, named in order of relative abundance. The total amount, however, is small. The obsidian has been scattered so widely over the island by man that one gets the impression of wider natural distribution than actually exists. The ratio and significance of the amount of obsidian will be discussed later.

The obsidian varies from a clear, greenish brown glass with a few perlitic areas, to a solid mass of spherulites with a little interstitial devitrified or crystallized glass. It has a conchoidal fracture except in some spherulite-rich types. The spherulites vary from microscopic size up to 4.0 mm. They are pale bluish gray when fresh and brownish gray when weathered. The spherulite-rich varieties have a ropy, fibrous structure with large vesicles, greatly elongated in the direction of flow. The index of the glass is $1.496 \pm .002$. The specific gravity varies from 2.333 to 2.398 depending upon the number of spherulites; the more abundant the spherulies, the higher the specific gravity. In thin sections the normal and most common type of obsidian, from which most of the spearheads or "mataa" were made, shows few, if any, phenocrysts. There are innumerable small spherulites in the glass, generally less than 0.1 mm. in diameter, and all arranged in planes parallel to the line of flow. Microlites of a greenish augite cloud the glass and cause a certain degree of opacity.

An abundant mineral in the glass is a triclinic feldspar, $n_\beta = 1.524 \pm .003$, $(+)2V = 80°$ to $90°$, $Z \wedge c = 8°$ to $10°$. Every crystal is a penetration twin, the twinning having apparently taken place on the orthodome with the side pinacoid (010) as the composition plane. This is the same type of triclinic feldspar that occurs in the Maunga Parehe olivine dacite tuff.

As the spherulites increase in number and size, there is apparently no change in the optical properties of the glass, although it does become darker and more waxy in luster. In view of the recent paper by Colony and Howard,[14] it is to be noted that nowhere have the spherulites deflected the flow lines of the pyroxene microlites. Such flow lines penetrate the spherulites freely and in undiminished amount, even in specimens consisting entirely of spherulites.

Analyses of the spherulites and the glass were made on one specimen. The glass made up 91 per cent of the rock area measurements and 89

[14] R. J. Colony and A. D. Howard: *Observations on spherulites*, Am. Mineral., vol. 19 (1934) p. 515-524.

per cent by volumetric determinations. An average of 90 per cent was assumed when calculating the actual composition of the rock, by combining the glass and the spherulites (Table 1, Nos. 1, 1a, and 1b). These analyses would indicate that, except for the marked difference in the ratio of soda to potash, there is practically no difference in composition between the spherulites and the glass. This would be expected if the spherulites contained much albite and quartz. An interesting feature of these analyses is the occurrence of ferrous iron in the spherulites and ferrous and ferric iron in almost equal quantities in the glass. Thin sections of the obsidian used for analyses showed a few scattered grains of zircon.

A variation of the Cerro Ourito obsidian was found in a small area on Rano Kao. This is a clear, dark greenish brown glass with a few visible phenocrysts of plagioclase and pyroxene. By use of heavy liquids it was possible to separate the various phenocrysts in the rock, and the following optical properties were obtained for the pyroxene and plagioclase: plagioclase, $n_\beta = 1.532 \pm .003$, (+); pyroxene, $n\alpha = 1.746$, $n^\beta = 1.757$, $n\gamma = 1.778 \pm .003$, (+)2V= 60°, $Z \wedge c = 50°$, $r > v$ — medium, X = green, Y = clear green, Z = yellowish brown. These data indicate almost pure albite and a pyroxene near aegirite. However, the optic angle, the birefringence, and the angle of extinction are not in close agreement with aegirite.

Origin of the Cerro Ourito Obsidian.—The writer estimated that there is somewhat less than 0.011 cubic miles of obsidian on the island. This, the maximum figure allowable, is probably too large but, as the volume of obsidian is important in what follows, the generous estimate is made. The area of the island is roughly about 55 square miles. On this basis it is probable that over 55 cubic miles of basalt have been extruded, and this amount of basalt has crystallized and differentiated to some extent during the crystallization. Thus, the amount of obsidian on the island is 0.02 per cent of the amount of basalt that has crystallized.

This small amount of obsidan could be formed from the basalt by several processes. There is no necessity to assume any process other than normal differentiation of the ocean basalt as it crystallized. The presence of olivine in every extrusive rock, almost up to the last flow, points to a relatively late and rapid change in the nature of the source magma. It would appear that if the magma was incorporating a less basaltic rock type in depth—for example, continental rocks—there would be a change manifest earlier in the period of volcanism, and olivine would have disappeared much earlier in the period of differentiation and extrusion.

The writer sees no great difficulty in deriving the obsidian from the magma which furnished the earlier volcanic rocks. In the rocks repre-

sented by analyses Nos. 3 and 8, Table 1, the differentiation had already proceeded well toward the obsidian. Considering the norms of the rocks and comparing analyses Nos. 3 and 1, it can be seen that by removing approximately 40 per cent of sodic andesine, 3½ per cent diopside, 9 per cent hypersthene, 5½ per cent ore, and 2 per cent quartz, a rock near obsidian in composition results. These are the minerals that would be expected to separate early from the andesite. To derive the rhyolite or obsidian from reaction of a basalt on a granite would require practically the melting down of the granite with very small admixture of basaltic material or very little differentiation.

COMPARISON BETWEEN ROCKS OF EASTER ISLAND AND THOSE OF OTHER PACIFIC ISLANDS

It is not easy to compare Easter Island with other islands in this general Pacific region. Definite data on the rocks of Sala y Gomez, the nearest island, are lacking. Isla Mas Afuera is the next nearest island, and no analyses are available of the rocks found there. Isla Mas a Tierra is a more basaltic island, as a whole, than Easter Island; an analysis of a nepheline basanite from there is quoted by Quensel.[15] The islands San Felix and San Ambrosia have been studied by Willis and Washington,[16] and analyses of two nepheline basanites are given in their paper on these islands; both rocks being from San Felix. The Hawaiian rocks tend to form alkaline types. The rocks from Hawaii, as described by Washington [17] and others, are not similar to those of Easter Island.

The writer sees no necessity for assuming that Easter Island was built up in any other way than by volcanoes rising from the depths of the Pacific. Archaeologists first advanced the idea that it was part of a great continental mass. They did not believe that the island would support a population sufficiently large to carve and move the statues that are found there. The continental idea became firmly rooted in the minds of island investigators; thus, Thomson and Routledge have reported rocks of continental origin or suggested their occurrence. Brown [18] needed to have a sunken continental mass to fulfill certain theories he had regarding Easter Island archaeology, and he used geology to prove the existence of such land masses. He calls the Rano Roraka tuff a conglomerate or "puddingstone with . . . angular fragments . . . that look like altered slate or fire-clay, and others that may be granite."

[15] P. D. Quensel: *Die Geologie der Juan Fernandez Inseln*, Geol. Instit. Upsala, Bull. 11 (1912) p. 263-264.

[16] Bailey Willis and H. S. Washington: *San Felix and San Ambrosia, their geology and petrology*, Geol. Soc. Am., Bull., vol. 35 (1924) p. 365-384.

[17] H. S. Washington: *Petrology of the Hawaiian Islands*, Am. Jour. Sci., 5th ser., vol. 6 (1923) p. 338-365.

[18] J. M. Brown: *The riddle of the Pacific* (1924). London.

Daly,[19] on the basis of these statements, included Easter Island in his list of islands where continental rocks were reported. Despite these several statements, there is no evidence on the island to support the idea of a sinking or sunken land mass about the island, as pointed out by Chubb.[20]

The occurrence of rocks of continental types has been reported and recorded from other mid-Pacific islands, but when specimens have been collected and examined petrologically the classification has often been found to be erroneous. For example, Lacroix[21] has proved that the "granite" of Borabora is an olivine gabbro which was no doubt intruded into the basalt. He has also shown the "leptynite" of Nukuhiva to be a trachyte of a type not common to the island.

The writer is of the firm belief that continental masses had no part in the formation of Easter Island, and the island stands as an argument against continental masses existing at depth in this part of the Pacific. There still remains, however, the question of the origin of the Albatross Plateau upon which Easter Island rests.

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[19] R. A. Daly: *Petrography of the Pacific Islands*, Geol. Soc. Am., Bull., vol. 27 (1916) p. 325-344.
[20] L. J. Chubb: *Geology of Galapagos, Cocos, and Easter islands*, B. P. Bishop Mus., Bull. 110 (1933).
[21] Alexander Lacroix: *La constitution lithologique des îles volcaniques de la Polynésie australe*, Acad. Sci. Paris, Mém., vol. 59 (1927) p. 27.

LLALLAGUA, BOLIVIA.
MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, JUNE 25, 1936.
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FIGURE 1. BLACK STAYTON BASALT
Showing hyaloöphitic texture, with pyroxene (augite) and labradorite phenocrysts in glass rendered opaque by magnetite dendrites. (S36) Plain light, X 33.

FIGURE 2. GRAY STAYTON BASALT
Characteristic loosely ophitic texture, with a tendency toward porphyritic texture. Compare with Figure 1. (S28) Plain light, X 33.

FIGURE 3. EAST END OF RIDGE SOUTH OF ELK LAKE
Thin hypersthene-olivine basalt flows of Battle Ax series overlying massive altered Sardine lavas. Prominent talus marks contact.

BASALTS

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PETROLOGY OF LATER TERTIARY AND QUATERNARY ROCKS OF THE NORTH-CENTRAL CASCADE MOUNTAINS IN OREGON, WITH NOTES ON SIMILAR ROCKS IN WESTERN NEVADA

BY T. P. THAYER

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INTRODUCTION AND ACKNOWLEDGMENTS

The Cascade Mountains in Oregon have been known since pioneer days as the central part of a volcanic mountain chain extending southward into California as far as Lassen Peak, and northward across Washington to Mt. Baker. The rugged nature of the mountains, the heavy

timber, and the volcanic rocks combined to render early geological reconnaissance difficult and inconclusive in many respects. It was soon recognized that the Cascade lavas overlie pre-Tertiary rocks in Washington and California, and that the conspicuous peaks are pre-glacial in age, and may have erupted within the last two or three centuries. However, it was by no means known, nor is it yet known, how many geological epochs are represented by the volcanic rocks of the Cascade Mountains, nor how many types of rocks occur, and in what proportions.

Hague and Iddings in 1883 gave the first detailed petrographic description of lavas from the younger peaks. They stated:

"These four great cones (Lassen Peak, Mt. Shasta, Mt. Hood, and Mt. Rainier) which may be taken as typical of the chain, are all andesite volcanoes, with extrusions of basalt breaking out upon their slopes."[1]

The chemical analyses which they quote are of andesite and dacite. They noted the gradational relation between basalts and andesites, and the reciprocal relation between olivine and hypersthene. Tridymite in andesite from Lassen Peak was also mentioned.

The second great contribution to the petrology of the Oregon Cascade lavas was the monograph by Diller and Patton on Crater Lake.[2] Like Hague and Iddings' work, however, the monograph was limited to younger lavas of the range—namely, those of Mt. Mazama. These two investigations, with other shorter contributions, led to the belief that the Cascade Mountains in Oregon were a rather simple accumulation of andesitic lavas surmounted by young volcanoes of similar general composition, whose lavas ranged from basalt to rhyolite.

I. C. Russell, in 1897,[3] was perhaps the first to suggest that part of the Cascade lavas had been folded, and that the younger volcanics lay unconformably on the older ones. Later papers on the Cascades substantiated Russell's observations, but contributed comparatively little to the petrology of the region. Hodge's accounts of the geology of Mt. Jefferson[4] and the Three Sisters[5] contain petrographic descriptions of olivine-oligoclasites and mugearites, among other lava types, but the discussions are primarily structural and physiographic in nature. His more recent work on the Columbia River canyon is not directly concerned with the petrology of the Cascade lavas, although they figure prominently in the argument.[6]

[1] Arnold Hague and J. P. Iddings: *Notes on the volcanoes of northern California, Oregon, and Washington Territory*, Am. Jour. Sci., 3d ser., vol. 26 (1883) p. 224.

[2] J. S. Diller and H. B. Patton: *Geology and petrography of Crater Lake National Park*, U. S. Geol. Surv., Prof. Paper 3 (1902) 167 pages.

[3] I. C. Russell: *Volcanoes of North America* (1897) p. 233-239. Macmillan Company, New York.

[4] E. T. Hodge: *Geology of Mount Jefferson (Oregon)*, Mazama, Ann. Ser., vol. 7, no. 2 (1925) p. 25-58.

[5] E. T. Hodge: *Mount Multnomah, ancient ancestor of the Three Sisters*, Univ. Oreg. Pub., vol. 3 (1925) p. 1-160.

[6] E. T. Hodge: *Columbia River fault*, Geol. Soc. Am., Bull., vol. 42 (1931) p. 923-934.

To date, the best comprehensive report on the Oregon Cascades is that by Callaghan, published in 1933. He briefly discussed the structure and stratigraphy of the range, which he divided into the Western Cascades and High Cascades on the basis of a marked unconformity in the lava sequence. The Western Cascade lavas—the older series—have been folded, faulted, and locally mineralized, and are composed of two chief lava groups.

> "One of these is characterized by black lavas with associated red agglomerates, tuffs, and rhyolite; the other of irregular flows of dominantly andesitic lavas with rhyolite and abundant andesitic and rhyolitic pyroclastic rocks. . . ."[7]

Callaghan says the High Cascades "are characterized by a large mass of gray olivine basalt, large cones composed mainly of hypersthene andesite, rhyolite and pyroclastics, and small cones, some of them quite recent, of andesite, calcic andesite, basalt, and possibly other types."[8]

The detailed structural and petrological description of Mt. Thielsen by Howel Williams, also published in 1933,[9] supplements Callaghan's paper. According to Williams, the main bulk of Mt. Thielsen is composed of gray basaltic lavas capped by a tuff cone, and Mount Bailey and Howlock Mountain are built of similar basalts. He also described intrusive plugs of coarse-grained basaltic types. Whereas Callaghan dealt chiefly with the Western Cascade, or older, lavas, Williams described in detail High Cascade lavas which, judging from degree of glacial erosion, are older than those of Mt. Mazama. It is significant that Callaghan and Williams, working independently, found great amounts of basaltic lavas in the High Cascade volcanic sequence.

While working in the vicinity of Mt. Jefferson, the writer found great thicknesses of gray basalts; these could be separated into distinct stratigraphic units, and therefore presented an opportunity for detailed petrographic study to determine, if possible, the local magmatic history. The fine-grained, in many places glassy, character of the rocks makes accurate determination of lava types very difficult, if not impossible, by microscopic means alone. The writer is therefore greatly indebted to the Geological Society of America for the generous grant from the Penrose Fund which enabled him to have 15 chemical analyses made. The analyses have made possible certain interpretations and conclusions which otherwise would have remained in doubt.

Permission of the United States Geological Survey to publish the original material on the Lousetown volcanics of western Nevada is greatly appreciated. The author also takes this opportunity to thank Dr.

[7] Eugene Callaghan: *Some features of the volcanic sequence in the Cascade Range in Oregon,* Am. Geophys. Union, Tr. (1933) p. 243.

[8] *Ibid.,* p. 249.

[9] Howel Williams: *Mount Thielsen, a dissected Cascade volcano,* Univ. Calif. Pub., Bull. Dept. Geol. Sci., vol. 23 (1933) p. 195-214.

Eugene Callaghan, Mr. Frank C. Calkins, Mr. W. D. Johnston, and Dr. Ian Campbell for their many suggestions and critical reading of the manuscript. The specimens on which this study is based are in the collections of the California Institute of Technology, where much of the laboratory work was done.

As the main Western Cascade (Miocene) sequence has been studied in detail by Buddington and Callaghan, it is described only in a general way in this paper. This investigation is concerned with the microscopical petrography and petrology of part of the Miocene lavas which have not been described previously, and of all the younger lavas in the area mapped except those of Mt. Jefferson. It is hoped that this paper may supplement the preceding articles to give a more complete picture of the petrology and petrography of the volcanic rocks of the Cascade Mountains in Oregon.

The region studied lies between that part of the Willamette River extending a few miles south of Salem and the summit of the Cascade Mountains in the vicinity of Mt. Jefferson and Olallie Butte. The best-exposed section is along the North Santiam River, about 50 miles south of the Columbia River and roughly parallel to it. The structure of the region, having been described in a recent publication,[10] will be considered in this paper only insofar as seems necessary for an understanding of petrological problems.

RÉSUMÉ OF STRUCTURE AND STRATIGRAPHY

In the North Santiam River district the Cascade Mountains are readily divisible into two parts, the Western Cascades and the High Cascades, on the basis of a marked unconformity. The volcanic rocks of the Western Cascades have been folded, faulted, locally intruded by dioritic plugs, and mineralized. The High Cascade volcanics are very fresh and apparently retain their initial attitudes. The time interval represented by the unconformity probably includes the last of the Miocene and possibly the earlier half of the Pliocene, although evidence on the chronology of the Cascade volcanics is admittedly scanty.

The only definitely dated formation in this part of the Cascade Mountains is the Middle Oligocene marine Illahe formation, which is exposed in the Salem Hills and the western margin of the Western Cascades. The Illahe formation grades eastward into equivalent continental deposits, the Mehama volcanics. Unconformably overlying the Oligocene formations are the basaltic Stayton lavas, which, though deeply eroded in places, are about 400 feet thick in the Salem Hills and thicken

[10] T. P. Thayer: *Structure of the North Santiam River section of the Cascade Mountains in Oregon*, Jour. Geol., vol. 44 (1936) p. 701-716.

toward the northeast. The Fern Ridge series of conglomerates and pyroclastics, ranging up to 1500 feet thick, are conformable on the Stayton lavas. This three-fold sequence is easily traced as far east as the crest

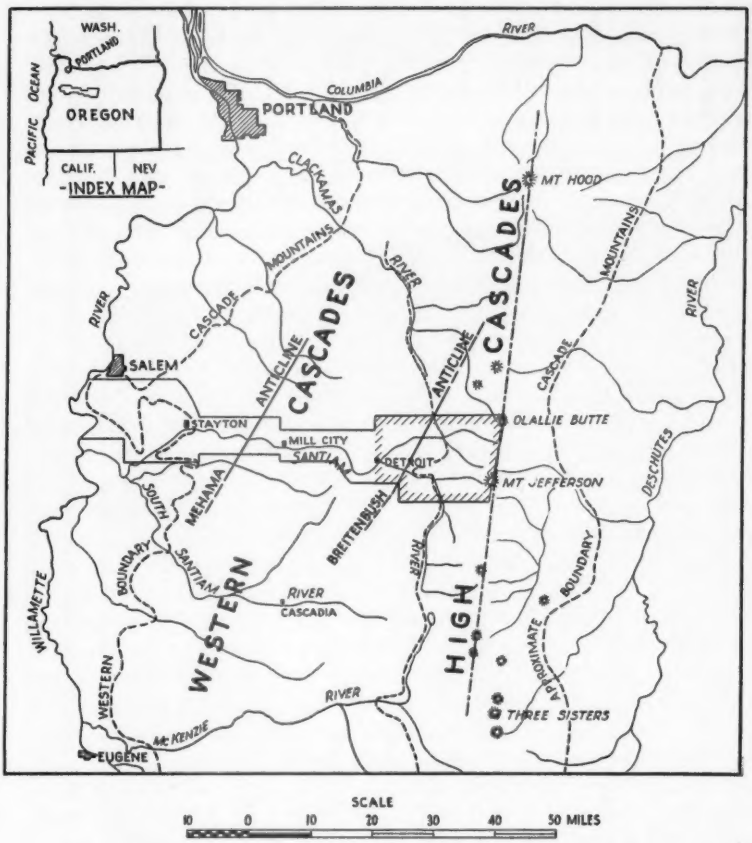
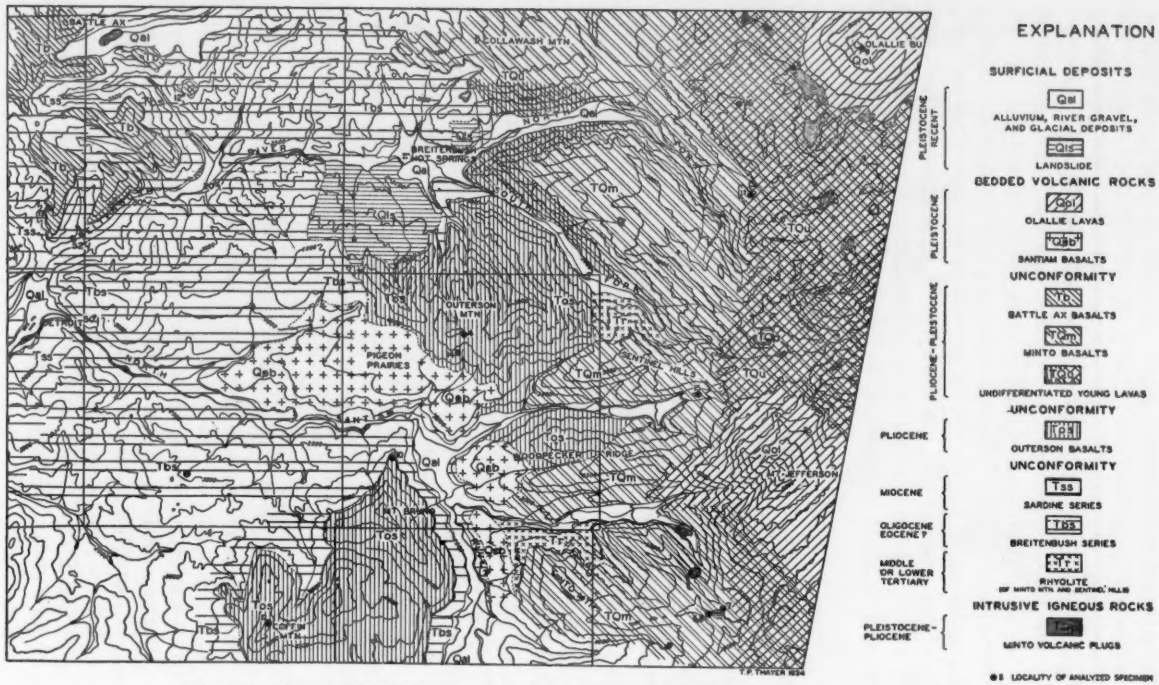


FIGURE 1.—*Northern portion of Cascade Mountains in Oregon*

of the Mehama anticline, about midway between Stayton and Mill City (Fig. 1); farther east lava flows become so numerous in the Mehama volcanics and Fern Ridge series that formational identity is lost. The three formations are therefore grouped in the Sardine series, which has an aggregate thickness of about 6000 feet and consists mainly of lavas. In the west limb of the Breitenbush anticline the lavas grade downward



GEOLOGY OF PORTIONS OF THE MILL CITY AND MT. JEFFERSON QUADRANGLES, OREGON.

FIGURE 2.—*Geology of portions of the Mill City and Mt. Jefferson quadrangles, Oregon*

into the Breitenbush formation, composed of water-worked and land-laid pyroclastics with some lavas, which at the base of the exposed section are rhyolitic. About 4 miles southwest of Detroit the Sardine lavas have been intruded by a body of diorite porphyry about a mile in diameter.

The Stayton lavas are considered the equivalent of the Middle Miocene Columbia River basalts. They resemble the latter in general character, and their thickness, as it increases toward the northeast, approaches that of the Columbia River basalt exposed in the Mehama anticline where the Clackamas River crosses it. The Stayton lavas also interfinger with and might be considered part of the Sardine series, which is probably equivalent to the Columbia River basalts, for the basalts were traced to the southern edge of the Estacada quadrangle by Barnes and Butler,[11] and the writer mapped the Sardine lavas within 5 miles of the same point. The regional structure shows such continuity that the two lava series must be stratigraphically equivalent with interfingering flows. However, to the writer's knowledge, the relation of the Columbia River basalts to the Sardine lavas and similar andesites in the Sandy River valley southwest of Mt. Hood has never been positively determined. The andesites overlying the Columbia River basalts in the region of the Columbia River gorge are High Cascade post-Miocene lavas.

In brief, the tentative correlations of the lavas in this section of the Western Cascades are as follows: The Stayton and Sardine lavas are equivalent to the Columbia River basalts and interfinger with them. The rhyolitic lavas at the base of the Breitenbush formation are probably John Day or Clarno in age.

The High Cascade volcanic rocks consist of at least four structurally distinct series separated by pronounced erosional unconformities. In chronological order they are: The Outerson series, the Minto lavas, and two probably nearly equivalent series, the Olallie lavas and the Santiam basalts. The Battle Ax series comprises lavas of probable Minto age which were erupted locally in the Western Cascades (Pl. 1, fig. 3). Discovery of similar High Cascade lavas within the physiographic boundaries of the Western Cascades is to be expected as field mapping progresses.

The bulk of the Outerson series consists of basaltic lavas with minor tuffs and lapilli beds but includes steeply dipping, bedded agglomerates which are abundant mainly in the lower portions. It was poured out from vents along the eastern scarp of the Western Cascades, and the lavas and breccias consequently dip in general eastward, their inclination usually being between 10 and 20 degrees. Some of the lavas which

[11] F. F. Barnes and J. W. Butler, Jr.: *The structure and stratigraphy of the Columbia River gorge and Cascade Mountains in the vicinity of Mt. Hood*, unpublished master's thesis, University of Oregon (1930).

lapped over onto the Western Cascade upland areas lie nearly horizontal, as in Coffin Mountain (Fig. 2). One of the most active eruptive centers of Outerson time was located a mile or two east of the summit of Outerson Mountain, where there is a great complex of intrusive plugs, coarse breccias, and steeply inclined flows. The contact relations of the Outerson and the Minto lavas indicate that the former were greatly eroded before the extrusion of the latter.

The Minto lavas formed shield volcanoes whose vents were near the axis of the present High Cascades. Three plugs of these volcanoes are exposed in the vicinity of Mt. Jefferson (Fig. 2). Dips of the lavas range from 6 or 8 degrees near the plugs to 1 or 2 degrees near the peripheries of the old cones. A radius of at least 11 miles is indicated for the cone whose center was 3 miles north of Mt. Jefferson and whose lower slopes extended northwestward beyond Collawash Mountain. Where the Outerson accumulations were high the Minto lavas flowed around them. This relation is well shown on a small scale in Woodpecker Ridge, where Woodpecker Creek flows in the trough between the southeast-dipping Outerson lavas and agglomerates and the west-dipping Minto lavas. The course of the North Santiam River valley south of the right-angle elbow probably originally followed the line of intersection of the Outerson eroded surface and the Minto initial surface. Where the Outerson deposits were thinner, the Minto lavas overrode them and lapped over on the Western Cascades, as in Collawash Mountain.

The broad basaltic cones of the Minto series were deeply dissected before the extrusion of the Olallie lavas. In the easternmost peak of the Sentinel Hills a thick andesitic flow from Mt. Jefferson dips 28 degrees northeast where it plunged into a valley cut in the Minto lavas, which there dip about 3 degrees westward. The Olallie lavas formed smaller and steeper cones than the Minto flows. Dips of 28 degrees were observed in thin black basalt flows on the west side of Mt. Jefferson, whereas the steepest dip seen in the Minto series was 8 degrees. Olallie Butte is a slightly dissected cone typical of many that were built up on the deeply eroded Minto series. The structural relations between the Olallie and the Minto lavas appear to be analogous to those between the tuff cone and the shield volcano which together form Mt. Thielsen.[12]

The Battle Ax lavas, identical in character with the Minto lavas, were poured out from a vent in the vicinity of Battle Ax Mountain, about 7 miles north of Detroit. The flows have been so deeply eroded that they are found now only as thick broad cappings on ridges. The close petro-

[12] Howel Williams: *op. cit.*, p. 201.

graphic similarity of the Minto and Battle Ax lavas, and the deep dissection of both formations, indicate that they are contemporaneous.

The Santiam basalts are possibly the youngest lavas in the area covered in this study, and they show the least variation. The individual flows are for the most part thin and lie nearly flat. The valley cut by the North Santiam River in the Minto and Outerson lavas was later filled by the Santiam basalts to a known depth of 1600 feet. The source of the flows appears to have been south of Outerson Mountain and was probably a fissure, as there is no evidence of a volcanic cone. The lavas flowed west down the North Santiam River valley nearly to Detroit and south up the valley beyond the southwestern corner of Minto Mountain (Fig. 2). The width of the valley filled by basalts indicates a considerable time break between the Minto and the Santiam series. The 1600-foot gorge in the Santiam basalts was cut in pre-Wisconsin time; hence the lavas are probably little, if at all, younger than the Olallie lavas, which are also pre-Wisconsin.

Correlation of the High Cascade lavas of the Mt. Jefferson region with similar lavas of the Mt. Hood and Columbia Gorge region is at present uncertain. The Outerson series, with its abundant pyroclastics, may be equivalent to the Dalles or Rhododendron formations, which Hodge considers Pleistocene or late Pliocene in age.[13] The Cascade formation probably includes lavas of both Minto and Olallie series; the earlier Intracanyon lavas may be equivalent to the Santiam basalts.

PETROGRAPHY

GENERAL STATEMENT

As the lavas and intrusive rocks of the Western Cascade sequence have been studied in some detail by Callaghan and Buddington,[14] the present article will be limited essentially to the Stayton lavas and the High Cascade volcanic rocks. General description of the Sardine lavas is included for purposes of comparison.

STAYTON LAVAS

The Stayton lava series in the outcrops west of Stayton consists entirely of basalts. East of Stayton it includes andesitic lavas, but these are subordinate west of the crest of the Mehama anticline; east of the anticlinal crest the basalts give way to labradorite andesites[15] of the Sardine series. The basalt flows range from medium gray to black.

[13] E. T. Hodge: *Columbia River Fault*, Geol. Soc. Am., Bull., vol. 42 (1931) p. 962.

[14] A. F. Buddington and Eugene Callaghan: *Dioritic intrusive rocks and contact metamorphism in the Cascade Range in Oregon*, Am. Jour. Sci., 5th ser., vol. 31 (1936) p. 421-449; *Metalliferous mineral deposits of the Cascade Range in Oregon*, U. S. Geol. Surv., Bull. (1937). In preparation.

[15] Eugene Callaghan: *Some features of the volcanic sequence in the Cascade Range in Oregon*, Am. Geophys. Union, Tr. (1933) p. 245-246.

They are locally uneven in thickness, but where exposures are good individual flows can be traced for 3 or 4 miles. Because of the regularity of the flows and prevailing dark color, the lavas closely resemble the Columbia River basalts. Both columnar and block jointing are common in the basalts, whereas the andesitic flows are commonly platy. Breccias and agglomerates form only a small part of the total thickness of the Stayton series.

The dark-gray to black basalts are usually slightly porous or vesicular. As seen in hand specimen, their texture varies from glassy to finely porphyritic. In the porphyritic types the feldspar phenocrysts are small and olivine phenocrysts may be present. In thin sections the texture is generally hyaloöphitic, with 20 to 30 per cent tabular feldspars and 15 to 25 per cent granular pyroxenes set in a glassy base, which constitutes 50 to 65 per cent of the rock. The feldspar is labradorite; it shows very slight or no zoning and is as much as a centimeter long, the average length being about 0.1 millimeter. The pyroxene is monoclinic and occurs normally as finely granular or rod-like masses scattered through the glass base, although in some cases it tends to form aggregates. Such aggregates are usually associated with olivine and may result from reaction between olivine and glass, as reaction rims of fine granular augite around olivine grains are common in olivine-bearing rocks.

Olivine is present in most of the dark lavas. In many flows it is recognizable megascopically as minute yellow grains. Under the microscope large olivines are rare, and much of the mineral occurs in small scattered granules with parallel optic orientation, associated with granular pyroxene, for which it is easily mistaken. Olivine is not abundant; at most it constitutes only a small fraction of one per cent of the whole rock. The fact that this readily altered mineral is almost invariably fresh except in weathered specimens bears witness to the lack of regional alteration of the Stayton lavas.

The dark color of these rocks is due to the minute crystals in the glass base. Most of the magnetite occurs as dendrites rendering the glass opaque under low magnifications (Pl. 1, fig. 1). Magnetite (?) dust and crystallites of pyroxene and feldspar add to the opacity of the groundmass.

The paler-gray Stayton lavas are usually porous, some being markedly so. The texture in hand specimens is finely porphyritic or even grained, the component minerals being recognizable under a hand lens. In one or two instances the lavas have a mottled appearance due to feldspar aggregates.

Under the microscope the gray lavas are characterized by diabasic texture, with 40 to 60 per cent tabular plagioclase, 20 to 30 per cent granular augite, occasional olivine, and 10 to 35 per cent glass or matrix (Pl. 1, fig. 2). The average size of feldspar is 0.3 millimeter, and of pyroxene and olivine about 0.1 millimeter.

The composition of the feldspar averages about andesine-labradorite (An_{50}). The larger phenocrysts of plagioclase are labradorite (An_{60}) and show the effects of resorption. Most of the feldspar is zoned from labradorite (An_{60}) to oligoclase (An_{15}). Some grains of clear oligoclase are interstitial to other feldspars. Inclusions are common in the plagioclase phenocrysts.

Monoclinic pyroxene and olivine are the only ferromagnesian minerals present. Augite occurs in granular aggregates molded on feldspar. Olivine is not uncommon (0.5 per cent) as small, partially resorbed remnants associated with augite. No reaction rims of fine granular pyroxene are developed, but the close association of augite and corroded olivine grains, the latter often showing parallel optic orientation, leaves little doubt regarding the origin of much of the augite. The olivine is the ordinary chrysolite variety, being optically negative with 2V nearly 90 degrees. No hypersthene was identified in the Stayton basalts.

The interstitial material in most flows of the pale type consists of colorless or very light-brown glass containing microlites of pyroxene and apatite. In one section (S17) most of the matrix is a micrographic intergrowth of oligoclase and pyroxene. The feldspar portion of many of the intergrowths is optically continuous with the oligoclase borders of adjoining plagioclases. A few long pyroxene needles are included in the intergrowth, which forms about 35 per cent of the slide. The rock contains no olivine and may be an andesite. Magnetite is scattered through the slides, chiefly as subhedral rods and grains, and forms about 3 to 4 per cent of most types.

Cristobalite occurs as small scattered masses in openings (Pl. 5, fig. 3). The mineral is complexly twinned in the pseudohexagonal manner common to tridymite, although measurement of refractive indices indicates that no tridymite is present. Some wedge-shaped twin segments appear to be strictly isotropic and may be metacristobalite. The clear-cut contacts between this mineral and other constituents of the rock indicate that it is not a zeolite, although some secondary minerals are present. The cristobalite in the gray Stayton basalts is apparently similar in occurrence to that in the San Juan basalts.[16]

[16] E. S. Larsen, John Irving, F. A. Gonyer, and E. S. Larsen, 3d: *Petrologic results of a study of the minerals from the Tertiary volcanic rocks of the San Juan region, Colorado,* Am. Mineral., vol. 21 (1936) p. 689.

TABLE 1.—*Chemical analyses and norms of*

	1.	2.	3.	4.	5.	6.	7.	8.
				Chemical analyses				
SiO_2	51.44	53.54	49.85	51.51	52.96	53.00	54.09	53.01
Al_2O_3	13.29	14.10	15.20	17.52	18.37	18.31	18.92	18.98
Fe_2O_3	2.16	1.73	1.69	1.51	2.72	2.37	1.43	1.93
FeO	12.00	10.33	10.50	6.69	5.78	5.97	5.97	5.53
MgO	4.28	4.59	6.25	6.74	5.38	5.30	5.33	5.12
CaO	8.28	8.54	9.43	8.78	8.03	8.11	8.29	8.96
Na_2O	2.93	3.14	3.08	3.34	3.82	3.92	3.83	3.72
K_2O	1.40	1.21	0.97	0.77	0.78	0.82	0.71	0.66
H_2O+	0.66	0.26	0.24	0.49	0.33	0.27	0.03	0.53
H_2O-	0.07	0.03	.....	0.15	0.16	0.14	0.07	0.20
TiO_2	2.80	2.01	2.50	1.12	1.25	1.26	1.04	1.11
P_2O_5	0.61	0.35	0.40	0.31	0.34	0.37	0.24	0.20
MnO	0.23	0.22	0.16	0.15	0.14	0.14	0.13	0.13
	100.15	100.05	100.27	100.08	100.06	99.98	100.08	100.08
				Norms*				
	III.5.3.4.	III.5.3.4.	III.5.5.(4–5)	II.5.4.(4–5)	II.5.3.4.	II.5.3.4.	II.5.4.(4–5)	II.5.4.(4–5)
Q	3.24	3.84	**	0.12	1.86	1.32	2.10	1.50
or	8.34	7.23	6.12	4.45	5.00	5.00	3.89	3.89
ab	24.63	26.72	26.20	28.30	31.96	33.01	32.49	31.44
an	18.90	20.57	24.46	30.58	30.58	29.75	32.25	33.08
wo	7.77	8.24	8.24	5.80	3.13	3.25	2.90	4.41
en	10.70	11.50	10.20	16.90	13.50	13.30	13.30	12.80
fs	15.97	14.52	9.24	9.50	6.73	7.13	8.32	7.00
mt	3.25	2.55	2.55	2.09	3.94	3.48	2.09	2.73
il	5.32	3.80	4.71	2.13	2.28	2.43	1.98	2.13
ap	1.34	0.84	1.01	0.67	0.67	1.01	0.67	0.34
hm	.....	.....	.....	.....	.....	.....	.....	.....

1. (836) Black Stayton basalt, 4 miles north of Stayton. A. Willman, analyst.
2. (828) Gray Stayton basalt. East of Oregon State Training School. A. Willman, analyst.
3. Olivine basalt. Cupola Rock, Lost Creek, Lane County. J. G. Fairchild, analyst.
4. (8122) Olivine basalt, Santiam type. Near summit Outerson Mountain. R. B. Ellestad, analyst.
5. (8131) Minto olivine basalt. Divide southwest of Hunts Cove. A. Willman, analyst.
6. (8130) Olivinitic micronorite, intermediate facies. Divide southwest of Hunts Cove. R. B. Ellestad, analyst.
7. (8129) Cristobalite micronorite, coarse facies of Minto plug. Southwest of Hunts Cove. R. B. Ellestad, analyst.
8. (8148) Olivine basalt, black-and-white type. Head of Olallie Lake. A. Willman, analyst.

* Norms calculated according to method of T. F. W. Barth: *Proposed change in calculations of norms of rocks*, Min. Pet. Mitt., vol. 42 (1931) p. 1–7.

** fo 3.78, fa. 3.67.

Sideritic carbonate showing dichroism in shades of yellow is present in some specimens, where it is plainly a product of weathering. Incipient weathering in many cases also produces much brown chloritic material which fills all porous spaces in the affected portion. The result is a very dark, almost black rind ranging up to 4 inches thick and strongly resembling the black basalts. Only by breaking through this outer layer can the pale color of the gray basalts be seen.

Chemical analyses (Table 1, analyses 1 and 2) indicate that the black basalts and the paler types of the Stayton lavas do not differ greatly in composition, except for a 2 per cent difference in total iron oxide. The color of each rock is determined, not by the proportion of mafic and felsic minerals, but by the distribution of magnetite in the glass base. Hence, the use of color as a means of differentiating between andesites and basalts in this region must be tempered with caution.

Nevertheless, the andesites of the Stayton series contrast strongly with the basalts in texture and structure. They are decidedly porphy-

Stayton lava and High Cascade volcanic rocks

9.	10.	11.	12.	13.	14.	15.	16.
			Chemical analyses				
53.50	53.54	53.74	54.45	57.86	58.89	60.11	66.93
18.51	17.81	19.13	17.92	18.83	15.76	18.20	16.32
5.56	3.48	1.41	3.02	1.71	5.42	1.62	3.22
2.54	4.59	6.03	5.05	4.07	2.47	3.64	0.76
4.78	4.53	5.17	4.27	4.19	2.39	3.02	1.39
8.93	7.93	8.46	7.49	7.23	5.51	5.99	3.73
3.78	3.39	3.68	3.68	3.92	4.65	4.37	4.59
0.62	0.91	0.69	1.18	0.74	1.58	0.97	1.74
0.22	0.90	0.21	0.52	0.20	0.56	0.74	0.39
0.07	0.83	0.08	0.36	0.17	0.54	0.18	0.13
1.14	1.15	1.00	1.18	0.83	1.55	0.71	0.56
0.18	0.50	0.27	0.50	0.12	0.55	0.14	0.20
0.12	0.14	0.12	0.22	0.10	0.15	0.09	0.09
99.95	99.70	99.99	99.84	99.97	100.02	99.78	100.05
			Norms				
II.5.4.(4–5)	II.5.4.(4–5)	II.5.4.(4–5)	II.5.3.4.	II.5.3.5.	II.4.3.4.	II.4.3.4.	I.4.3.4
5.94	7.20	2.28	6.03	9.42	12.90	12.18	24.00
3.34	5.56	3.89	7.23	4.45	9.45	6.12	10.01
31.96	28.82	31.44	31.44	33.01	39.30	37.20	38.77
31.69	30.58	33.36	28.36	31.41	17.51	26.97	17.79
4.99	2.20	2.78	2.20	1.51	2.55	0.81	†
12.00	11.30	12.90	10.70	10.50	5.80	7.60	3.50
.....	3.96	8.32	5.15	4.88	.....	4.09	.....
5.10	5.10	2.09	4.41	2.55	3.71	2.32	1.16
2.13	2.13	1.98	2.28	1.52	3.04	1.37	1.06
0.67	1.34	0.67	1.34	0.34	1.34	0.34	0.34
2.08	.....	.....	.....	.....	2.88	.....	2.40

9. (S151) Olivine basalt, Olallie series. Summit of Olallie Butte. A. Willman, analyst.
10. (S142) Olivine basalt, Outerson series. North end of Mt. Bruno. A. Willman, analyst.
11. (S163) Gray-and-white olivine basalt. Northwest of Pyramid Butte. R. B. Ellestad, analyst.
12. (S114) Coffin Mountain basalt, olivine-bearing. Summit of Coffin Mountain. A. Willman, analyst.
13. (S126) Hypersthene-cristobalite andesite. French Creek Ridge, north of Detroit. R. B. Ellestad analyst.
14. (S118) Glassy augite andesite, Outerson Mountain. A. Willman, analyst.
15. (S152) Hornblende-bearing pyroxene andesite. Knob between Jeff and Whitewater creeks at 5000 feet. A. Willman, analyst. II.4.3.4.
16. (S150) Hornblende-bearing pyroxene andesite. Summit of Double Peaks, west of Olallie Lake. A. Willman, analyst.

† Corundum 0.41.

ritic, with feldspar and pyroxene phenocrysts set in an aphanitic to glassy, dark groundmass.

In thin section two generations of feldspar are apparent, the phenocrysts ranging from 0.5 millimeter to 1.5 millimeters or larger, and the groundmass plagioclase ranging from 0.02 millimeter to sub-microscopic dimensions. Flow structure is usually well developed. The phenocryst plagioclase is zoned from andesine-labradorite to oligoclase, and the groundmass feldspars are oligoclase. The pyroxene is mainly augite in medium to small grains with some semi-prismatic crystals. Rounded hypersthenes also occur as phenocrysts. The groundmass, whether felted or glassy, contains abundant microlites of feldspar and pyroxene and powdery magnetite.

SARDINE LAVAS

The lavas of the Sardine series are dark gray, coarsely porphyritic to aphanitic. They differ from the Stayton lavas in that ophitic textures are very rare, either glomeroporphyritic or flow textures predominating.

Feldspar phenocrysts average one or two millimeters in length but rarely reach a centimeter.

The phenocryst feldspars are zoned; they normally have cores of sodic bytownite or calcic labradorite and rims that range from calcic andesine to oligoclase. The groundmass feldspar varies from sodic labradorite in the more common basaltic types to andesine or oligoclase in the less common dacitic types. Plagioclase constitutes about 75 per cent by volume of most sections studied. The primary ferromagnesian minerals are augite, hypersthene, and olivine. Augite is the dominant phenocryst pyroxene whereas hypersthene occurs sparingly, especially in the upper part of the lava series. In a few slides hypersthene prisms are surrounded by augite occurring not as small granules, but as single continuous crystals enclosing only the prism and side pinacoid faces of the hypersthene. Olivine is represented in a few flows by pseudomorphs of chloritic material or carbonate and quartz surrounded by granular augite reaction rims. The groundmass pyroxene is nearly all granular or prismatic augite. The pyroxenes, both phenocryst and groundmass, are largely represented by chloritic pseudomorphs. Primary quartz phenocrysts occur in a few flows.

The groundmass varies from clear or pale-brown glass to felted or intergranular mixtures of feldspar, pyroxene, and magnetite, and variable amounts of glass. Commonly devitrification has taken place, and where the alteration accompanying diorite intrusion was extensive the glass base has been completely crystallized.

The predominance of feldspar over pyroxene in the Sardine lavas contrasts with the high percentage of mafic constituents in the Stayton basalts. This contrast is graphically shown by a comparison of the content of alumina and iron oxide in the two series as shown in Figure 3. The basic types of Stayton lavas are basalts whether classified according to femic-salic ratio of the rock or the soda-lime ratio of the feldspar. If the Sardine lavas are classified on the former basis with the limits ordinarily accepted, they are all andesites; according to the composition of the feldspars, the more calcic types are basalts. For the latter rocks, which are calcic andesites or low-pyroxene basalts, Callaghan has used the descriptive term labradorite andesite.[17]

HIGH CASCADE LAVAS

General Statement.—The High Cascade lavas as a whole are fresh, pale- to dark-gray, and more or less porous. In the area studied, they range in composition from basalts to dacites and exhibit great textural variation.

[17] Eugene Callaghan: *op. cit.*, p. 246.

The distinction between basalts and andesites in the High Cascade sequence is necessarily somewhat arbitrary. The logical factor to use in classification of the two types of rocks in a series low in ferromagnesian constituents, as most of the High Cascade lavas are, is the composition of the feldspar. However, where the feldspars show such strong zoning as is common in these lavas, determination of their average composition, even with the aid of chemical analyses, is very difficult if not impossible.

In the absence of more exact and usable criteria, the presence or absence of olivine has been found to be a fairly reliable index to the basaltic or andesitic nature of these lavas. All the olivine-bearing flows analyzed for this project contain less than 55 per cent SiO_2, which figure is in the zone ordinarily regarded as on the border between andesite and basalt. The more silicic lavas, besides being olivine-free, commonly show the marked textural and mineralogical variations characteristic of andesitic lavas. In view of these considerations, and, for convenience in petrographic description, all olivine-bearing High Cascade lavas will be termed basalts, and other varieties will be grouped in the andesite category unless other evidence points to the contrary. On this basis the following chief types of lavas are recognized: Olivine basalts, hypersthene basalts, pyroxene andesites, pyroxene-cristobalite andesites, and pyroxene-hornblende andesites.

Basalts.—The basalts in this portion of the High Cascades are represented by several varieties, differing greatly in texture, color, and other characters. The contrast between some types both in the hand specimen and under the microscope is so striking that their nearly identical chemical composition is rather surprising.

The basaltic lavas may be divided into two groups on the basis of the presence or absence of hypersthene. These may in turn be subdivided as follows:

Hypersthene-free basalts

SANTIAM TYPE: Pale-gray porous lavas, fine- to medium-grained, olivine abundant.

COFFIN MOUNTAIN TYPE: Pale-gray, very fine-grained rocks with excellent flow structure and platy fracture. Olivines rare and very small.

Hypersthene-bearing basalts

MINTO TYPE: Light-gray to blue-gray, finely porphyritic, more or less porous lava containing small reddish olivines. These lavas grade into pale noritic plug rocks.

BLACK-AND-WHITE BASALTS: Fine-grained, very porous, dark-gray or black lavas with minute olivines and small white feldspars. Usually contain hypersthene.

MASSIVE GRAY-AND-WHITE BASALTS: Dense and coarsely porphyritic lavas with large white feldspar glomerocrysts and small green or yellow olivines in a fine-grained gray groundmass.

DARK-GRAY OUTERSON TYPE: Rather massive, finely porphyritic basalt containing much serpentinized olivine.

GLASSY BLACK HYPERSTHENE BASALTS: Olivine-free.

The Santiam type of basalt is normally very porous, medium- to pale-gray lava characterized by abundant yellow olivine phenocrysts. The flows, as typified by the Santiam series, average 15 to 20 feet in thickness and are massive or obscurely columnar.

Specimen S122, which represents a thin flow and feeder dike 10 feet wide just below the summit of Outerson Mountain, is the coarsest variety of the Santiam type. Analysis 4 of Table 1 represents this rock. Phenocrysts of olivine, labradorite, and monoclinic pyroxene averaging 5 millimeters to 6 millimeters in length, form 32 per cent of the thin section. In the groundmass, feldspar with random orientation averages 0.05 millimeter to 0.06 millimeter. Augite prisms and globulites form 20 per cent of the matrix, and plagioclase and brown glass about 40 per cent each.

The feldspar is calcic labradorite (An_{60} to An_{75}) zoned sharply at the extreme edge, and partly rimmed with oligoclase (An_{15}). Magnetite grains occur in olivine, but are most abundant in the glass base. The rock contains approximately 50 per cent by volume plagioclase, 6 per cent olivine, 16 per cent monoclinic pyroxene, 2.5 per cent magnetite, and 25 per cent glass.

Slide S106, a pale aphanitic basalt with small olivine phenocrysts, is more typical of the basalts of the Santiam series. Corroded grains and poikilitic plates of olivine amounting to 6 per cent of the slide are set in a groundmass of tabular feldspar, granular pyroxene, and magnetite. The fabric of the matrix varies from loosely ophitic to obscurely intergranular. The plagioclase is labradorite zoned sharply to oligoclase (An_{15}) on the margins. Monoclinic pyroxene constitutes 15 or 20 per cent of the rock. A small amount of light-brown glass is present. Cristobalite occurs in vesicles as very rare small aggregates.

One of the most deceiving rocks, in hand specimen, of the entire High Cascade sequence is the Coffin Mountain basalt, the medium-gray platy Outerson lava which forms the summit of Coffin Mountain. It is aphanitic with a micaceous sheen given by minute cleavage faces on parallel tabular feldspars. The flow texture seen in thin section is exceptionally perfect for a rock of basaltic composition. Only a few small olivines attaining 0.5 millimeter in size disturb the rippled parallelism of the 0.03 millimeter- to 0.3 millimeter-long labradorite laths, many of which are somewhat bent (Pl. 4, fig. 1). The olivines are more or less surrounded by granular augite, obviously a reaction product. Granular

augite and colorless glass are interstitial to the feldspars. Analysis 12 of Table 1 is of this rock.

In several ways this lava is in marked contrast to the other High Cascade basalts. The well-developed flow structure and uniform grain size are unusual in a series of prevailingly porphyritic and random textures. The flow is at least 500 feet thick; whereas the average for other basalt flows is between 10 and 50 feet. Apparently the Coffin Mountain lava issued as a viscous flow in which crystallization was nearly complete; this is inferred from the texture and is confirmed by the fact that, in spite of its great thickness, the flow has been recognized only in Coffin Mountain.

The basalts which derive their name from Minto Mountain (Fig. 2), in the east end of which their relation to noritic plugs is well exposed, constitute the bulk of the Minto and Battle Ax series and are represented in the Outerson series. They are probably the commonest lava type in this portion of the High Cascades.

In hand specimen they vary from porous, pale varieties with millimeter-long feldspars and small olivine phenocrysts to bluish-gray, rather dense, platy types resembling andesite except for the presence of small reddish olivines. Columnar structure is rarely developed, and then in crude fashion. Good platy fracture is not unusual. Flows are normally thin, averaging 25 feet or less. In a thousand-foot section exposed on the south side of Collawash Mountain the flows average 10 feet or less and usually overlie an equal or greater thickness of agglomerate or flow breccia. The more porous varieties are indistinguishable in the field from the finer of the Santiam hypersthene-free basalts; hence it is impossible to make an accurate estimate of the relative abundance of hypersthene-bearing and hypersthene-free flows of this type.

The micro-texture ranges from hyaloöphitic or glomeroporphyritic to porphyritic intergranular. Olivine phenocrysts range in diameter from less than 0.5 millimeter to 1 millimeter, and occupy 2.5 to 4 per cent of the slides. Plagioclase glomerocrysts attain lengths of 5 millimeters, averaging 1 millimeter or less. Pyroxene phenocrysts approximate the olivines in size, and may be augite or hypersthene or both. Brief descriptions of typical slides will best give a conception of the mineralogic and textural variations.

A thin section of a typical medium-gray, somewhat porous basalt, which would be taken for a Santiam-type flow, reveals feldspar glomerocrysts as much as 5 millimeters across and small grains of olivine and pyroxene in a fine-grained random-textured groundmass. The feldspar phenocrysts show intense zoning, much of it oscillatory, from labradorite to oligoclase, the latter forming extremely narrow rims. The cores of

the large phenocrysts were resorbed to a great extent and replaced by zoned, more sodic plagioclase. Small rounded olivine grains with granular augite rims form 2 per cent of the slide. The pyroxene phenocrysts are larger equant augites and ragged hypersthene prisms, some of which appear to be monoclinic.[18] The groundmass consists of minute tabular feldspars which are zoned from andesine to oligoclase (An_{15}), interstitial oligoclase, granular augite, and magnetite. Apatite is commonly enclosed in the groundmass feldspars. Small interstitial masses of cristobalite occur sparingly.

The hypersthene-bearing type of Minto basalt, which shows more andesitic tendencies in that it is commonly platy and less granular and shows better flow structure, is well exemplified by S131 (Table 1, analysis 5). This lava grades imperceptibly into the intrusive norite (Table 1, analyses 6 and 7).

In hand specimen the lava is pale bluish gray, with small reddish olivines in an aphanitic groundmass. The micro-texture is seriate porphyritic, with phenocrysts of plagioclase, olivine, augite, and hypersthene (Pl. 3, fig. 2). Plagioclase laths range from 1.4 millimeters in length to almost submicroscopic size, the average minimum length being 0.08 millimeter. The groundmass is intergranular, grading into patches characterized by an extremely fine mosaic of feldspar and poikilitic hypersthene.

The plagioclase phenocrysts are labradorite (An_{60}) with narrow oligoclase (An_{15}) rims. The larger grains are raggedly scalloped by resorption and contain inclusions of olivine, magnetite, and apatite. In the extremely fine mosaics the feldspar is untwinned and probably albitic. The olivine (3 per cent) is much resorbed and usually associated with minute augites or poikilitic hypersthene which may completely enclose kernels of olivine. Hypersthene, the dominant pyroxene, occurs as subhedral poikilitic plates, half a millimeter long in many places, molded on feldspar, and as very minute, extremely "wormeaten" plates in the mosaic patches. Small stout prisms of augite are evenly distributed in the groundmass except where it is accumulated as reaction rims around olivine. A few extremely ragged poikilitic augite plates were noted. Magnetite grains are scattered more or less uniformly through the groundmass, except where associated with olivine and pyroxene as a reaction product. Glass is rare or absent; quartz may be present in the fine mosaic patches, but was not positively identified.

The Minto volcanic plugs, from which Minto lavas were extravasated, are exposed in Park Butte north of Mt. Jefferson and in the east end

[18] J. Verhoogen: *Monoclinic "Hypersthene" from the Cascade lavas*, Am. Jour. Sci., 5th ser., vol. 33 (1937) p. 66.

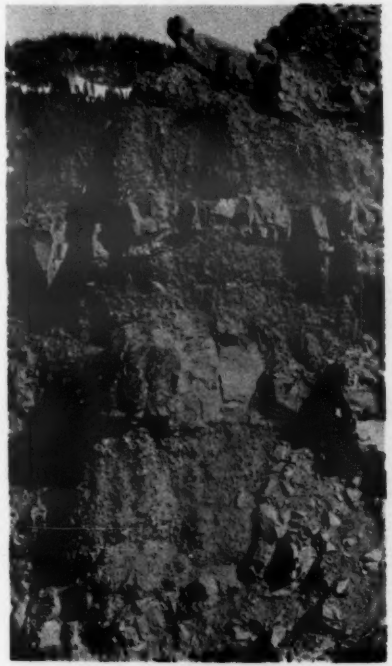


FIGURE 1. BASALT FLOWS IN EAST END OF MINTO MOUNTAIN
About a mile from an exposed plug. Thickest flow near the center is about 6 feet thick.

FIGURE 2. PYRAMID BUTTE
Hornblende-pyroxene andesite plug. Glaciated outcrop in foregound is massive gray-and-white basalt.

FIGURE 3. UPPER PORTION OF MINTO PLUG IN HUNTS COVE
Showing gradational relations of massive coarse plug rock (below) and fine-grained platy facies.

FLOWS AND PLUGS

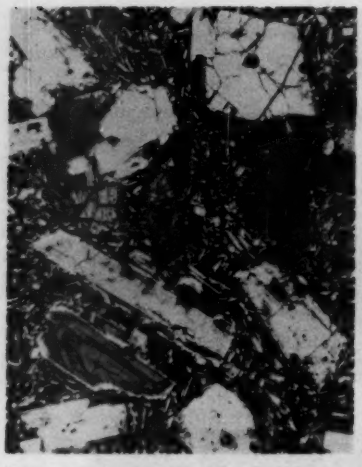


FIGURE 1. MASSIVE GRAY-AND-WHITE BASALT
Containing large olivine and plagioclase phenocrysts in groundmass of feldspar, pyroxene, magnetite, and glass. Oscillatory zoning of central portion and sharp zonation of margins of plagioclase phenocryst in lower left are typical, except that the crystal appears to have been broken. (S163) Crossed nicols, X 33.

FIGURE 2. MINTO OLIVINE BASALT
Grading into micro-norite. Comparison of this rock with figures 3 and 4 reveals the change in grain size and development of marked porphyritic texture in the lava. Prevalence of random texture is interpreted as evidence of a high degree of fluidity in the magma. (S131) Crossed nicols, X 33.

FIGURE 3. COARSE-GRAINED MICRO-NORITE OF MINTO PLUGS
Tabular calcic plagioclase, less-formed sodic plagioclase, poikilitic augite, and subprismatic hypersthene. (S129) Crossed nicols, X 33.

FIGURE 4. OLIVINE-BEARING INTERMEDIATE FACIES OF MINTO MICRO-NORITE PLUGS
Rounded olivines (white) and tabular labradorites in a finer-grained groundmass of plagioclase, augite, and hypersthene, the whole exhibiting random texture. (S130) Crossed nicols, X 33.

PHOTOMICROGRAPHS

of Minto Mountain. In the latter locality the gradational relations of micro-norite and lava are well exposed in the more southerly of the two plugs.

The intrusive rock grades from a light-colored rock containing white plagioclase as much as 2.5 millimeters in length and yellow ferromagnesian minerals of comparable size to fine-grained typical blue-gray olivine basalts. Intermediate textural facies are represented in such perfect gradation that no particular type can be justly termed a "border phase". The finer-grained facies are pale-gray to buff-colored granular rocks. The coarsest facies occur only as small irregular bodies suggestive of pegmatite in granite, and are limited to the central portion of the plug. As shown in figure 3 of Plate 2, the coarser plug rock is sparsely jointed and forms cliffs, whereas the finer facies near the contacts are usually platy, but less so than surface flows. No dikes of coarse rock were observed cutting lavas, although small pipe-like masses of the finer facies commonly intrude flows, some of which are vesicular. Careful search revealed no sharp intrusive contacts between plug and flow.

Under the microscope the coarse-grained facies appear as finely noritic, even-grained rocks with a variation in grain size from 0.2 to 2 millimeters (Pl. 3, figs. 3 and 4). Plagioclase forms about 70 per cent by weight of the rock, occurring as squarish plates and small laths zoned from labradorite (An_{60}) to albite (An_5). Albite with all indices below 1.54 occurs interstitially as distinct crystals, some of which show twinning. Most of the albite is untwinned, whether occurring as margins on twinned calcic crystals or as interstitial grains, and there is no reason to doubt its primary origin. Hypersthene is the dominant pyroxene, exceeding augite in the ratio of about 3 : 2. It occurs as medium-sized to small subhedral pleochroic prisms, the larger of which contain drop-like inclusions of plagioclase. Augite characteristically forms large poikilitic plates molded on calcic feldspar. Much of the hypersthene probably preceded augite in crystallization and some of it appears to be monoclinic. Rare corroded olivine grains are present. Apatite needles are sparingly scattered through the feldspar. Cristobalite occurs as interstitial fillings which show obscure twinning, apparently being optically uniaxial or isotropic.

Microscopically the intermediate phases are a textural and mineralogical compromise between the coarse plug facies and the lava. The texture is seriate porphyritic, with an older generation of olivine and plagioclase and a younger one of plagioclase, hypersthene, augite, and magnetite (Pl. 3, figs. 3 and 4). The range in grain size in any given slide is not so great as in the lava, but much greater than in the center of the plug. The plagioclase shows the same extreme zonation, with

discrete albite grains in the groundmass. Hypersthene is the dominant pyroxene, but not markedly so. It and augite occur as poikilitic plates molded on feldspar and associated with olivine and magnetite as reaction products. Subhedral prisms of hypersthene are also numerous. Most of the augite forms either small grains between feldspar laths or granular reaction rims on olivine. Deeply corroded irregular grains of olivine are as much as 0.7 millimeter long. Cristobalite is present as minute interstitial masses.

The mineralogic variation of the several facies may be accounted for by different rates of cooling and crystallization and by enrichment of the plug in silica by rising emanations. The rarity of olivine in the coarser facies may be considered the result of almost complete resorption during slow cooling, which permitted establishment of equilibrium between liquid and solid in the olivine-pyroxene discontinuous reaction series. Relative abundance of olivine in the peripheral plug facies and lavas indicates undersaturation of the magma with respect to silica or failure to establish equilibrium, which may be due to rapid cooling. The holocrystalline nature of the marginal facies and the small amount of quartz (1.32 per cent) in the norm suggest that the former is the more probable explanation. Analysis reveals the chemical identity of the lava and marginal facies, and indicates higher silica content in the coarse facies. The greater amount of cristobalite in the plug is in agreement with the chemical analysis in this regard.

The presence of cristobalite, rather than quartz, in a plug of this type supports Larsen's argument[19] that cristobalite is precipitated as an unstable form in the presence of an abundance of mineralizers. In a mass cooling slowly enough to allow development of fine-grained gabbroid textures, the temperature drop should be so gradual that inversion of unstable forms of silica would be appreciable if they were formed from residual glass while the rock was at a high temperature, as in the case of a dry melt. Nevertheless, no quartz was observed in the Minto plugs, although tridymite may be present. Crystallization of the rock was practically complete before the cristobalite was formed; this is shown by the texture of the rock and the fact that albite, at the acidic end of the plagioclase series, occurs as well-formed grains. It would be fortuitous if the residual glass had the exact composition of albite, and a small amount of free silica would normally be left. Any excess silica now appears as cristobalite. In addition, the chemical analyses suggest that silica was introduced into the plug rock. It seems very probable, therefore, that part of the cristobalite represents residual glass

[19] E. S. Larsen, *et al.*: *Petrologic results of a study of the minerals from the Tertiary volcanic rocks of the San Juan Region, Colorado,* Am. Mineral., vol. 21 (1936) p. 693.

crystallized under the influence of mineralizers, and that part, perhaps most, of it is a direct deposit from a gas.

The texture of the lavas clearly indicates that crystallization began in the conduit of the volcano but had not progressed far before extrusion and rapid cooling. The irregular form of the small intrusive pipes and their random texture prove that the magma was fluid when it entered the upper part of the conduit. Uniform grain size in a given specimen of the intrusive rock indicates cooling under uniform conditions, and the relatively large grain size in the coarser facies suggests slow crystallization. The superposition of the platy finer facies over the massive coarser type of plug rock demonstrates that the plug crystallized in the conduit some hundreds or possibly thousands of feet from the surface.

Williams has suggested that the Mt. Thielsen and Howlock Mountain plugs, which are petrographically nearly identical with the Minto plugs, were intruded as viscous masses with sufficient force to cause upturning of tuff beds, although he does not describe shearing along the contacts of the coarser rocks.[20] Absence of shearing points to a liquid rather than a highly viscous or nearly solid mass. The dark lavas against which the paler intrusives in many places have sharp contacts are similar in general character to the normal Minto lavas, and do not show appreciable effects of metamorphism. It seems possible that sharp contacts between plug rock and lavas originally gradational might be explained by stoping of a chilled border facies which was broken up by movements accompanying later eruptions. Removal of border phase in this way would permit still-fluid magma to come in contact with solidified lava which, however, was hot enough so that chilling of the magma would not occur. Where stoping occurred, well-defined intrusive contacts would appear between coarse facies and lava or border facies as the case might be; elsewhere the relations would be gradational, as is true in many places. Upturning of the tuffs is admittedly more easily explained by intrusion of a highly competent semi-solid plug of the Pêléean type, but might also be accomplished by transmission of pressure through a liquid mass. The apparently contradictory evidence of the upturned tuffs and the gradational contacts with lavas suggest that there is some as yet undiscerned factor in the mechanism of intrusion of the Minto and Thielsen plugs.

The so-called black-and-white basalts are dark-gray porous lavas. They are characterized by millimeter-long feldspar phenocrysts which, with small olivine grains, stand out conspicuously against the dark groundmass. Olallie Butte is almost entirely built of this type of flow,

[20] Howel Williams: *Mount Thielsen, a dissected Cascade volcano,* Univ. Calif. Pub., Bull. Dept. Geol. Sci., vol. 23 (1933) p. 206.

which is also common on Mt. Jefferson.[21] A score or more of these flows, averaging less than 20 feet thick and dipping 28 degrees to the west, are well exposed on the west side of Mt. Jefferson at an elevation of about 6000 feet.

This type of Olallie lava is marked by porphyritic random texture. The feldspars range from 3.5 millimeters in length to submicroscopic size, and the olivine phenocrysts are somewhat smaller. The plagioclase is strongly zoned, much of it in oscillatory fashion, from calcic labradorite (An_{70}) to andesine or (less commonly) oligoclase. Olivine phenocrysts constitute about 5 per cent or less of the slides studied, with appreciable variation in amount from flow to flow. All slides contain abundant monoclinic pyroxene as large plates, small subhedral prisms, and grains. Hypersthene occurs in most sections as well-formed prisms, some of which are jacketed by augite. Hypersthene, however, is not everywhere present. Olivine is more abundant in the hypersthene-free than in the hypersthene-bearing rocks. Magnetite is distributed through the slides as granular inclusions in feldspar, pyroxene, and olivine, and as octahedra in the glass base. The matrix consists of yellow to brown glass containing varying proportions of fine feldspar, pyroxene, and magnetite. In one slide minute cristobalite grains were noted on the walls of openings. The approximate mode of a typical hypersthene-free flow (S148) is as follows: plagioclase, 60 per cent by volume; olivine, 4.5 per cent; augite, 14.5 per cent; magnetite, 3 per cent; glass, 18 per cent.

The most massive flows of the High Cascade lavas are the coarsely porphyritic gray-and-white basalts. These flows are more than 100 feet thick, break into huge blocks without any semblance of columnar structure, and resist glacial abrasion in a manner similar to granite. These were noted only on the High Cascade plateau south and west of Olallie Butte and probably form part of the Olallie series.

This medium-gray lava is rather dense. White feldspars up to half a centimeter long and yellow or green olivines as much as 1.5 millimeters in diameter stand out against the aphanitic groundmass. Plagioclase occurs as glomerocrysts with cores of bytownite (An_{80}) zoned to labradorite or andesine, and as small tabular crystals of labradorite or andesine, the latter in some cases zoned to sodic oligoclase. Olivine forms between 3.5 and 6 per cent of the specimens (Pl. 3, fig. 1). Resorption phenomena are well shown in specimens which were not chilled too rapidly.

[21] E. T. Hodge: *Geology of Mount Jefferson, Oregon*, Mazama, Ann. Ser., vol. 7, no. 2 (1925) p. 45 and 46.

The quantitative relation between olivine and hypersthene differs somewhat from that observed in most cases. Usually as olivine decreases in amount hypersthene increases, but in one of these specimens containing 6 per cent olivine, the augite-hypersthene ratio is about 3:2, whereas in a slide containing only 3.5 per cent olivine no hypersthene was observed. This variation from the normal may be caused by uneven distribution of olivine, although this was not apparent in the hand specimen. Colorless or light-brown glass constitutes about 10 per cent of these rocks. Analysis II in Table 1 is of this lava type.

An interesting type of olivine basalt found only in the Outerson series is a dark-gray, finely porphyritic basalt containing abundant partially serpentinized olivine. The texture is porphyritic with phenocrysts of plagioclase, augite, and olivine in an intergranular groundmass (Pl. 4, fig. 2). The feldspars are zoned from bytownite (An_{80}) or labradorite to oligoclase (An_{20}). The oligoclase rims are narrow. The pyroxene is mainly granular augite in the groundmass, although both augite and hypersthene occur as phenocrysts. The augite shows much twinning and hour-glass structure. Phenocrysts of olivine and serpentine pseudomorphs after olivine are very abundant (6 per cent). The serpentinization is apparently a late magmatic effect, as the olivines in gray porous lavas above and below this flow are very fresh. A minute amount of clear brown glass is present.

The base of the Outerson series is marked by a few black glassy hypersthene basalts. They exhibit closely spaced irregular jointing which produces small blocks of the "brickbat" type.

The microscope reveals an unusual arrangement of tabular labradorite and prismatic pyroxene suspended in brown glass (Pl. 4, fig. 4). There are a few large phenocrysts of plagioclase with clear bytownite cores, intermediate "wormeaten" zones, and clear labradorite rims. The average length of the feldspars is about 0.15 millimeter. Hypersthene, with refractive indices indicating a content of 20 to 28 per cent $FeSiO_3$, is the dominant pyroxene, although augite is abundant both as prisms and as granules. Extremely thin borders of monoclinic pyroxene are developed on the prism faces of a few hypersthenes. Magnetite is scattered rather evenly as inclusions in feldspar and as grains in the glass base which has an index of refraction of about 1.52. The glass forms between 25 and 50 per cent of the slides examined, and as the proportion of glass decreases the number of fine cloudy inclusions increases rapidly.

Although this rock is olivine-free and for that reason, as previously explained, might be placed with the andesitic lavas of the High Cascades, it is classed with the basalts because of the calcic nature of the feldspar and the clear glass base. It may be said that in general the

basalts contain clear glass as a base whereas in andesites the matrix is a felted or granular mixture of pyroxene, feldspar, and glass.

Andesites.—The High Cascade andesites in the area studied vary considerably in composition, texture, and mineralogy, but they do not include so many distinct types as the basalts. The andesites would doubtless show more variety if the lavas of Mt. Jefferson had been included in this project, for Mt. Jefferson, like Mt. Mazama and the other pre-Wisconsin High Cascade peaks which are probably of Olallie age, is mainly composed of andesitic lavas, and it is in these younger large peaks that the greatest variation is found. This fact is amply demonstrated by the published studies in the Crater Lake [22] and Lassen Peak regions.[23]

These more acidic lavas do not fall into distinct types so readily as the basalts. The andesites are in general markedly porphyritic, quite porous, and light- to medium-gray or buff. For simplification of description they may be segregated into pyroxene types and hornblende-bearing pyroxene types. The pyroxene andesites may be further subdivided into (1) glomero-porphyritic types with hypersthene, (2) hypersthene-cristobalite andesite, and (3) glassy augite andesite.

Glomeroporphyritic pyroxene andesite, a medium-gray or purplish-gray porphyritic lava with white plagioclase glomerocrysts ranging up to 3.5 millimeters in length, and dark pyroxenes up to 1.5 millimeters long set in a more or less aphanitic groundmass, is probably the most abundant andesite. In some instances the pyroxene tends to form aggregates with the feldspar phenocrysts.

The plagioclase phenocrysts are usually zoned from labradorite (An_{60}) to oligoclase (An_{15}); the rims of the latter are narrow. The groundmass feldspar consists of laths averaging andesine in composition (An_{40}), and some interstitial oligoclase is usually present. Subhedral hypersthene is the dominant phenocryst pyroxene, although plates and prisms of clinopyroxene are common if not abundant (Pl. 4, fig. 3). The latter is more abundant in the groundmass, although hypersthene is also present. The ratio of hypersthene to clinopyroxene in the rock as a whole probably averages about 1 : 1.

The groundmass is normally an intimate mixture of fine feldspar laths, pyroxene prisms or grains, and granular magnetite, with oligoclase, high-temperature silica, and some clear glass as interstitial material. Cristobalite or tridymite are important constituents of the matrix. No quartz was noted in any of the many slides studied.

[22] J. S. Diller and H. B. Patton: *Geology and petrography of Crater Lake National Park*, U. S. Geol. Surv., Prof. Paper 3 (1902) 167 pages.

[23] Howel Williams: *Geology of the Lassen Volcanic National Park, California*, Univ. Calif. Pub. Bull. Dept. Geol. Sci., vol. 21, no. 8 (1932) p. 195-385.

Some of the paler and finer-grained andesites of this type are probably chemically close to dacites. In them the phenocryst plagioclase is zoned from andesine (An_{40}) to sodic oligoclase, and the calcic cores in some cases are partly replaced by albite. In one slide, where the groundmass feldspar is zoned, phenocrysts show the effects of resorption, and many of the smaller laths contain as inclusions numerous small clinopyroxene prisms. More commonly the groundmass feldspar is extremely fine-grained with indefinite outlines. It appears to be albite or oligoclase, with possibly some interstitial orthoclase. Most of the pyroxene occurs as fine prisms in the groundmass. Phenocrysts of hypersthene and augite are not uncommon, and the augite is apt to show resorption and marginal granulation. The hypersthene-clinopyroxene ratio varies widely, with the latter normally dominant.

The groundmass in the more silicic andesites is extremely finely granular and seems to be holocrystalline. Prisms of pyroxene and magnetite grains are readily recognized. Aside from these minerals, the matrix has a low refractive index and is rather weakly birefractive. There is undoubtedly much oligoclase or albite (An_{5-15}), and possibly some orthoclase, but most of the interstitial material appears to be cristobalite or tridymite, judging from its negative relief and weak birefringence. Although the rocks are obviously saturated types and appear to be holocrystalline, no quartz was positively identified. The only possible traces of hornblende observed were some aggregates of granular augite and magnetite which may have resulted from complete resorption of the amphiboles. Similar aggregates derived from hypersthene or hornblende were noted in the hypersthene-cristobalite andesite and are described in detail in this same section.

The greatest lava sheet individually traceable in this area is a hypersthene-cristobalite andesite flow which was extruded on the southeast flank of Battle Ax volcano. After much erosion, this flow averages about 300 feet in thickness and originally covered an area of at least 15 square miles; it underlies most of the area of Battle Ax lavas south and east of Elk Lake (Fig. 2). A dike of this same rock 150 feet thick cuts the Minto lavas about 4 miles southwest of the summit of Olallie Butte and 2 miles west of Breitenbush Lake. No smaller bodies of this lava type were noted.

Megascopically the rock is a pale-gray, porous lava containing conspicuous rounded hypersthene phenocrysts, ranging up to 3 millimeters in length, and dark elongate ferromagnesian "ghosts". In parts of the flow, as on the summit of Gold Butte, the hypersthenes are absent and the ghosts are abundant. Glistening flakes of cristobalite paramorphic after tridymite are numerous in the small cavities with scattered specularite crystals.

Thin sections reveal strong glomeroporphyritic texture, with feldspar and pyroxene glomerocrysts and granular augite-magnetite pseudomorphs in an intergranular groundmass of pyroxene, plagioclase, magnetite, and cristobalite. Average phenocrysts are about 0.75 millimeter long, groundmass grains approximate 0.08 millimeter. The plagioclase shows variation in composition from bytownite (An_{85}) to oligoclase (An_{18}). Oligoclase rims on phenocrysts are narrow. The groundmass feldspars are zoned from andesine to oligoclase, and much of the latter is also present as discrete grains.

Hypersthene, the dominant pyroxene, is present as large rounded phenocrysts seen megascopically and as minute prisms in the groundmass. Refractive indices indicate that the $FeSiO_3$ content of the phenocrysts is about 17 per cent, and that of the groundmass prisms about 28 per cent. The latter contain a great abundance of minute bubbles. The hypersthene-augite ratio is about 2 : 1 or 3 : 1.

Clinopyroxene occurs as small phenocrysts, as prisms in the groundmass, and in elongate granular aggregates with magnetite. These aggregates are the ghosts seen in hand specimen. The margins of the ghosts are usually magnetite-free and the pyroxene is coarser, indicating resorption of the iron ore and growth of augite (?). The end stage appears to be a loose aggregate of augite prisms with little or no magnetite. It is not certain what mineral these ghosts represent. Their elongate form suggets hornblende, although none was found. In a few places, small resorbed hypersthene grains were observed, which may be remnants of large phenocrysts. The absence of phenocryst hypersthene and the abundance of ghosts in the Gold Butte portion of the flow suggests that the ghosts may have been formed at the expense of the orthopyroxene. It seems significant that, although hypersthene phenocrysts commonly have been resorbed in part, no "smut" rims were noted on remnant grains. This suggests that the replaced mineral was probably hornblende. It is, moreover, unlikely that resorption of iron-poor hypersthene would be accompanied by precipitation of magnetite at the same time iron-rich hypersthene was crystallizing. The weight of evidence points to hornblende, now completely resorbed, as the original phenocryst mineral.

The sequence of crystallization of the ferromagnesian minerals in this lava appears to be (1) hornblende, (2) magnesian hypersthene and augite, (3) ferriferous hypersthene and clinopyroxene, and (4) magnetite. This apparently simple sequence is complicated by extensive resorption and re-precipitation, particularly of the iron. An unusual abundance of mineralizers, evidenced by the amount of cristobalite in the lava, may be partly accountable for the apparently erratic behavior of the iron.

It appears that some highly ferriferous mineral, probably hornblende, crystallized to such an extent that the fluid fraction of the magma was impoverished in iron. Separation of plagioclase also began very early. With falling temperature and shifts in equilibria represented by the change in crystallization from amphibole to pyroxene, magnesian hypersthene was formed. The marked break in grain size suggests that extravasation of the lava ended crystallization of the magnesian hypersthene. The ghosts show little or no evidence of disruption by movements of the magma, yet the groundmass crystals around them show good fluidal structure. This suggests that destruction of the hornblende occurred simultaneously with crystallization of the groundmass, during and after the extrusion of the lava. The escape of mineralizers probably rendered the amphibole unstable, and it broke down into augite and magnetite; the absence of unaltered cores and the granular nature of the two minerals imply that the process may have resembled the decomposition of hornblende in dry fusions more closely than resorption and reaction of the olivine-pyroxene type.

With destruction of the hornblende the liquid fraction was controlled by the equilibrium relations of pyroxene rather than amphibole, and was iron-poor; it accordingly resorbed much of the magnetite which had suddenly become available. In fact, so much magnetite was resorbed that the hypersthene phenocrysts apparently became unstable and were partially resorbed. With continued drop in temperature ferriferous hypersthene in equilibrium with the glass was formed. The presence of magnetite in the groundmass shows that more iron was resorbed than could be re-precipitated in pyroxene, and indicates that an excess of iron oxides must be present in a melt from which these minerals are crystallizing.

In their experiments on the system FeO-MgO-SiO_2, Bowen and Schairer found that hypersthene could contain as much as 85 per cent $FeSiO_3$.[24] In saturated melts with silica all FeO went into pyroxene, the final product being pyroxene, and tridymite. In this lava the final hypersthene contains only about 30 per cent $FeSiO_3$, yet was precipitated from a liquid containing excess iron oxide. The final product (disregarding plagioclase), instead of being hypersthene and tridymite, is hypersthene, magnetite, and cristobalite. It would seem that the chief effect of mineralizers and ferric oxide in a melt related to the FeO-MgO-SiO_2 system is to increase the MgO : FeO ratio in the pyroxene, and concentrate iron oxide in the residual glass. When the magnesia was exhausted the iron oxide perforce crystallized as magnetite.

[24] N. L. Bowen and J. F. Schairer: *The system, MgO-FeO-SiO_2*, Am. Jour. Sci., 5th ser., vol. 29 (1935) p. 151-217.

Cristobalite constitutes about 10 per cent of this lava (Pl. 5, fig. 4). It occurs as aggregates up to 0.75 millimeter long in cavities, in stringers up to 0.5 millimeter long, and as interstitial grains in the groundmass. The larger masses are commonly paramorphic after tridymite,[25] exhibiting pseudohexagonal penetration twinning, the segments having negative elongation. Curved fracture is characteristic, and cleavage is well shown in only a few instances. Birefraction is very weak, the range of indices being between 1.485 and 1.487. The mineral is not a zeolite, because it is infusible before the blowpipe. Moreover, the norm of the rock (Table 1, analysis 13) contains 9.42 per cent quartz which is unaccounted for if the mineral is not a form of free silica, as the rock contains little if any glass. The mineral is not an alteration product, because all contacts with other minerals are sharp, and all minerals in the thin sections are absolutely fresh. Careful index tests with immersion oils failed to reveal any tridymite in the rock.

Tridymite but not cristobalite has been described by Larsen [26] as an essential mineral in the groundmass of siliceous andesitic and rhyolitic lavas. The tridymite-bearing San Juan lavas described by him and shown in his figures 2 and 3 closely resemble the more silicic High Cascade andesites in that they have dense glassy groundmasses; the latite shown in his figure 4, except for finer texture, has a closer resemblance to the holocrystalline cristobalite andesite. The feldspar-cristobalite relation in the groundmass is similar to that in spherulites. Is it possible that conditions in a spherulite could be simulated in a massive lava flow—in this case more than 300 feet thick and extending over several square miles—rich in mineralizers, so that cristobalite instead of tridymite formed in the groundmass? A large amount of volatile constituents is indicated by the abundance of cristobalite in cavities of the lava; unusual conditions are suggested by the large number of paramorphs after tridymite. There is no evidence that the silica was introduced from an outside source after solidification of the lava.

The only apparent difference between the Battle Ax lava and the similar Minto dike is the total absence of visible rounded hypersthenes, smaller size of the magnetite-augite ghosts, and smaller porous spaces in the dike. Cristobalite aggregates in the dike are smaller and more evenly scattered as a result of the smaller vesicles. Mineralogically and texturally the two rocks are almost identical.

Glassy augite andesites were noted only in the upper part of the Outerson series. The type specimen from Outerson Mountain, an anal-

[25] A. F. Rogers: *The occurrence of cristobalite in California*, Am. Jour. Sci., 4th ser., vol. 45 (1918) p. 225-226.

[26] E. S. Larsen, *et al.*: *Petrologic results of a study of the minerals from the Tertiary volcanic rocks of the San Juan region, Colorado*, Am. Mineral., vol. 21 (1936) p. 687-689.

FIGURE 1. COFFIN MOUNTAIN TYPE OF OLIVINE BASALT
Showing fluidal texture and unusual uniformity of grain size. Hematite-stained olivine appears black. (S114) Plain light, X 33.

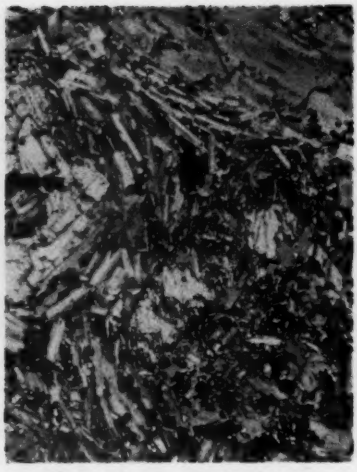


FIGURE 2. DARK-GRAY OUTERSON BASALT
Containing partially serpentinized olivine (white) near center of field, augite, and calcic labradorite in intergranular groundmass of plagioclase, clino-pyroxene, magnetite, and glass. (S120) Crossed nicols, X 80.

FIGURE 3. PYROXENE ANDESITE OF MINTO SERIES
Prominent hypersthene and labradorite phenocrysts lie in a matrix of sodic plagioclase, clino-pyroxene, magnetite, and glass. (S157) Plain light, X 33.

FIGURE 4. GLASSY HYPERSTHENE BASALT
From base of Outerson series. Tabular labradorite and prismatic pyroxene, mainly hypersthene, are suspended in clear brown glass with granular magnetite. (S138) Plain light, X 120.

PHOTOMICROGRAPHS

FIGURE 1. HORNBLENDE-BEARING PYROXENE ANDESITE

Magnetite-rimmed hornblende phenocrysts contrast with sharply bounded augite crystal at bottom of field and well-formed plagioclase phenocrysts. Groundmass consists of sodic feldspar, pyroxene, magnetite, glass, and high-temperature forms of silica. (S152) Plain light, X 33.

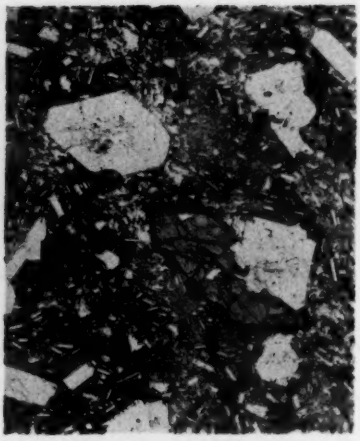


FIGURE 2. HORNBLENDE IN HORNBLENDE-BEARING PYROXENE ANDESITE

Has no magnetite rim against feldspar with which it is interlocked. Groundmass of section is plagioclase, pyroxene, magnetite, high-temperature silica, and glass. (S150) Plain light, X 33.

FIGURE 3. CRISTOBALITE IN GRAY OLIVINE BASALT OF STAYTON SERIES

Note freshness of feldspars and sharp contacts with cristobalite, which shows strong negative relief. (S28) Plain light, X 240.

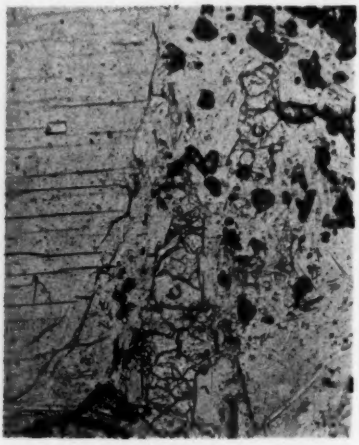


FIGURE 4. CRISTOBALITE (c) IN HYPERSTHENE-CRISTOBALITE ANDESITE.

Situated in intergranular matrix of sodic plagioclase and pyroxene (mainly hypersthene). Note freshness of labradorite phenocryst in left portion of field, and sharp zoning at margin indicated by changes in relief. (S141) Plain light, X 120.

PHOTOMICROGRAPHS

ysis of which is shown in Table 1, is an aphanitic, dark, smoky-gray rock with minute feldspar spots. In thin section it is characterized by hyalopilitic texture with a few phenocrysts of augite and andesine-labradorite. Weak fluidal structure is indicated by parallel arrangement of many of the feldspar crystallites, which are too small for accurate determination as to composition. The glass base, constituting 95 per cent of the slide, is clouded with fine magnetite and dust. The Boca Cave flow, from a vent 2 miles southeast of Outerson Mountain, is much more glassy, appearing vitreous in hand specimen. In addition to occasional pyroxene, feldspar, and hornblende phenocrysts, it contains fragments of several varieties of finely porphyritic augite andesite. The feldspar is sharply zoned, much resorbed bytownite. The augite crystals are perfectly formed. One twinned crystal of green hornblende with a magnetite reaction rim was noted in thin section. The glass base (98 per cent) is nearly opaque with magnetite dust and crystallites and exhibits good flow structure.

Hornblende-bearing pyroxene andesites seem to be restricted geographically to the area southwest of Olallie Butte and east of the North Fork of the Breitenbush River (Fig. 2). The individual flows were limited in extent, probably to less than a mile in most instances, and appear to have been viscous when first extruded, as the lavas are porous and in some places almost frothy. Some of the andesites are unquestionably flows intercalated with Minto pyroxene andesites. Others, of which that in Pyramid Butte (Pl. 2, fig. 2) is a good example, probably were intruded as viscous masses of the plug dome type. Glacial scour has removed all evidence of any cinder cones which may have been formed by explosive action.

These lavas are porous, pale- to medium-gray, buff, or reddish rocks exhibiting plagioclase and ferromagnesian phenocrysts in hand specimen. The hornblende crystals are usually less than 5 millimeters long, although, on the north side of the Double Peaks, crystals up to 2 inches long were found.

Most of these andesites (Table 1, analyses 15, 16) contain plagioclase phenocrysts zoned from bytownite (An_{85}) or labradorite to oligoclase (An_{15}). The zoning is commonly oscillatory, and the calcic cores are much smaller than in the pyroxene andesites. The groundmass feldspar is mainly andesine, with some well-crystallized oligoclase, although most of the latter is interstitial. Prismatic hypersthene is the dominant phenocryst pyroxene, the crystals being scattered singly or aggregated as glomerocrysts. Augite prisms with granular resorbed borders also occur as phenocrysts. Hornblende is present as phenocrysts only (Pl. 5, figs. 1 and 2). It is the common variety with maximum extinction angles

of 16 degrees in sections cut parallel to the C axis. The prisms are moderately stout, and without exception have exsolution rims of fine magnetite at contacts with the groundmass; at contacts with other phenocrysts the rim is usually absent (Pl. 5, figs. 1 and 2).

The groundmass is a felted mixture of feldspar, pyroxene, high-temperature silica, magnetite, and glass. In some sections the glass is the dominant element of the groundmass.

A small inclusion of olivine-bearing basalt was noted in one slide. The borders of the inclusion were vague, and the olivine shows the effects of strong corrosion.

The Pyramid Butte plug appears to be more dacitic and differs in several ways from the average hornblende andesites. Inclusions of fine black hornblendic rock are common, and in rare cases exceed 2 feet across. The inclusions are porous, and are practically pure hornblende, the crystals averaging about 5 millimeters in length. The plagioclase phenocrysts are zoned from calcic andesine to oligoclase (An_{20}), the cores showing effects of corrosion. In many instances they are poikilitic and contain inclusions of glass and dusty dark matter. Strong oscillatory zoning is the rule. Strongly pleochroic hypersthene is the dominant pyroxene. The edges of the hypersthene phenocrysts are apt to be ragged, although the abundant slender crystals in the groundmass are well formed. Augite is sparingly present as poikilitic phenocrysts with granular margins, but is abundant in the groundmass as stout prisms from half to one-tenth the length of the hypersthenes, which average about 0.25 millimeter in length.

The hornblende is pleochroic in shades of yellow-green and brown with maximum extinction angles ($Z' \wedge c$) of 16 degrees. Many crystals are beautifully sharp without signs of resorption or magnetite reaction rims. Some are surrounded by heavy magnetite rims; others have developed borders of granular augite. The end-product of reaction appears to be magnetite and granular or raggedly prismatic augite reminiscent of the aggregates in the hypersthene-cristobalite andesite.

The groundmass of the rock is composed of tabular feldspars, pyroxene prisms, magnetite, a large proportion of cristobalite or tridymite, and a small amount of brown glass. Flow structure is well developed.

PETROLOGY

GENERAL STATEMENT

The generally accepted view has been that Cascade lavas are mainly andesites associated with subordinate basalt and rhyolite. This is certainly true for the Western Cascade lavas, but probably not for the High Cascade lavas in Oregon. Any general discussion of the petrology

of the Cascade magmatic province must take into consideration the break at the end of the Miocene (?), which is apparently magmatic as well as structural. Much more detailed field work also is needed in the Cascade region in Oregon and Washington for complete unraveling of the volcanic history.

WESTERN CASCADE LAVAS

Petrographically and chemically the Stayton lavas, as represented in the region west of Stayton, are definitely basaltic. The prevalence of ophitic texture, widespread olivine, and great lateral extent of individual flows suggest their close relation to plateau basalts. Both the dark and the paler gray basalts are only slightly, if actually, oversaturated with respect to silica. The salic-femic ratio is between 1.4 : 1 and 1.25 : 1, values typical of silicic basalts. The gray lavas have the higher ratio and higher silica content.

In view of the structural relations of the Stayton lavas and the Sardine lavas, which have already been discussed, the andesitic flows of the former series may be considered marginal flows of the latter group of volcanics. The Sardine lavas range from olivine basalt to rhyolite, but are dominantly andesitic. The outstanding character of both the andesites and basalts is the low content of ferromagnesian minerals. The salic-femic ratio in the norm exceeds 2 : 1 even in the most basic lava analyzed. The feldspathic character of the lavas is reflected in their high lime and alumina content (Fig. 3). The more basic types may be called either low-pyroxene basalts or calcic andesites. These are the lavas for which Callaghan used the name labradorite-andesite, as the feldspar is labradorite or more calcic plagioclase in both the norm and the mode.

Lateral variation of the Western Cascade lavas is more evident than vertical variation. In the North Santiam River region dacitic and rhyolitic flows appear mainly in the upper part of the Miocene section, although dacites occur below the middle of the Sardine Mountain section. Lateral variation in composition is shown in the relation of the Sardine lavas and the Stayton and Columbia River basalts. It is also demonstrated in southern Oregon where basalts, which in the Rogue River valley resemble Columbia River lavas, apparently interfinger with andesite and rhyolite flows in the South Umpqua River district.[27] The vertical variation in the North Santiam River region suggests progressive differentiation of a single magma, whereas the lateral variation, apparently fully as great, indicates simultaneous eruption from several magmas of very different composition, or from differentiated layers of the

[27] Eugene Callaghan: *Some features of the volcanic sequence on the Cascade Range in Oregon*, Am. Geophys. Union, Tr. (1933) p. 244.

same magma. Intercalation of rhyolite between basalt flows, such as was noted east of Ashland, on the highway to Klamath Falls, suggests magmatic conditions similar to those in the Paulina Mountains discussed by Williams.[28]

HIGH CASCADE LAVAS

Basaltic lavas range stratigraphically from the base to the top of the High Cascade volcanic succession. Basalts constitute the entire Santiam series of flows, are the predominant lava type in the Outerson and Ollalie series (excepting possibly Mt. Jefferson in the latter series), and are probably dominant over andesites in the Minto series. It should be remembered that for convenience all olivine-bearing rocks are classified in this paper as basalts, hence most of the High Cascade lavas contain olivine. Most olivine-free rocks of andesitic appearance are therefore classed as andesites, although they may actually be more basaltic than andesitic. The basalt-andesite ratio is more likely to be underestimated than overestimated.

Abundance of basaltic lavas in the High Cascade sequence is not peculiar to the North Santiam River district. Williams has indicated that the main bulk of Mounts Thielsen and Bailey and all Howlock Mountain are composed of olivine-bearing basalts.[29] All the intracanyon lavas found in the Columbia River gorge and tributaries, the McKenzie River valley, many tributary valleys of the Willamette River, and the valleys of the North Umpqua and Rouge rivers, are gray olivine basalts similar to those making up the Santiam series. The Santiam basalts are really intracanyon or valley flows. There is unquestionably much variation in the age of the different valley flows. In a reconnaissance of the Cascade Mountains in Oregon during the summer of 1931, in which the writer assisted Callaghan, gray olivine-bearing basalts of pre-intracanyon age were found in great abundance from the Ashland-Klamath Falls highway, near the California State line, to the Cascade summit near Clackamas Meadows, 20 miles south of Mt. Hood.

Young olivine basalts as defined in this paper are abundant in the region north and west of Mt. Hood. These rocks have been generally referred to as the "Cascade andesites" by previous workers; it is the writer's belief that this name is a misnomer in that it places undue emphasis on andesitic lavas in a series which is definitely, though perhaps not strongly, basaltic. The writer has examined only a few thin sections of rocks from the Columbia River region, but those seen were identical with lavas from the North Santiam district. In hand speci-

[28] Howel Williams: *Newberry Volcano of central Oregon*, Geol. Soc. Am., Bull., vol. 46 (1935) p. 300-303.

[29] Howel Williams: *Mount Thielsen, a dissected Cascade volcano*, Univ. Calif. Pub., Bull. Dept. Geol. Sci., vol. 23 (1933) p. 201.

mens taken along the Sandy River east of Troutdale, from Larch and Pepper mountains a few miles farther east, and from Trout Creek Hill in the Wind River valley northeast of Bonneville, small to conspicuous olivine phenocrysts are present in variable amounts. Underwood Mountain, on the north bank of the Columbia River nearly opposite the town of Hood River, is a basaltic cone possibly of Minto or Outerson age.

By far the greater number of the scattered cones surmounting the Cascade plateau north of Olallie Butte are relatively low broad structures. Larch Mountain, 27 miles east of Portland, and Peavine Butte in the upper Clackamas River basin are good examples. Although apparently slightly eroded, their surface slopes are low; the west side of Larch Mountain slopes only 1700 feet in 2 miles, or about 9 degrees, and slopes on three sides of Mt. Defiance, a similar peak west of the Hood River valley, average about 7 degrees. (*See* the Mt. Hood and Vicinity topographic map for location of Larch Mountain and Mt. Defiance, and the Hood River sheet for location of Underwood Mountain.) These cones are in reality small steep shield volcanoes and must have been built up by fluid basaltic lavas. In the Clackamas River watershed the low cones dominate the topography whereas in the vicinity of the Columbia River gorge more rapid erosion has destroyed many; hence in the latter locality the abundance of young basaltic lava is by no means proportional to the number of well-preserved cones, but is much greater. The northern limit of these basaltic cones and lavas is apparently in the Mt. Adams region of southern Washington, as gray basalts are not reported in the vicinity of Mt. Rainier.[30] These cones are pre-Wisconsin, yet well preserved as a group, hence are tentatively correlated with the Olallie series.

Overemphasis has apparently been placed on andesites in the High Cascades in Oregon because of lack of detailed field and petrographic work in areas beyond the slopes of the large peaks. The Crater Lake monograph [31] is a good case in point, as Mt. Mazama is typical of the young pre-glacial peaks (Olallie series) of the High Cascades. No "shield" of basaltic lava was described, probably because the older lavas are not exposed within the limits of the Park. Hence the rocks described in the monograph should not be regarded as representative of High Cascade lavas as a whole, but as those of the best known High Cascade volcano. In Mt. Thielsen the younger tuff cone is so small that the shield volcano is the dominant structure as regards volume.

[30] H. A. Coombs: *The geology of Mount Rainier National Park*, Univ. Wash. Pub., Bull. Dept. Geol. Sci., vol. 3 (1936) p. 175-188.

[31] J. S. Diller and H. B. Patton: *Geology and petrography of Crater Lake National Park*, U. S. Geol. Surv., Prof. Paper 3 (1902) 167 pages.

Perhaps more emphasis should be placed on study of the deeply incised flanks of the High Cascades than on the summit regions, where thick sections are rarely exposed except in the high peaks, which, after all, form only a rather small fraction of the bulk of the whole range.

High Cascade lavas in the vicinity of Mt. Jefferson range from true basalts to siliceous dacitic rocks. All are characterized by notably high alumina content which is reflected in a high salic-femic ratio. In lavas containing as little as 52.5 per cent SiO_2 the ratio is 2.25 : 1, being even higher than in Western Cascade lavas of comparable basicity. Most of the basalts are probably just saturated or slightly oversaturated with respect to silica, although the Cupola Rock lava (Table 1, analysis 3) is unsaturated.

Basalts of similar silica content exhibit considerable difference in amounts of other oxides. For instance, two olivine-bearing augite basalts differing only 0.5 per cent in SiO_2 differ 1.46 per cent in total Al_2O_3. One, S122, is from the Outerson series; the other, S148, is an Olallie lava. Both are porphyritic. The more olivinitic flow (6 per cent olivine as compared with 4.5 per cent) also contains 1.62 per cent more MgO. It is evident that if these basalts represent a differentiate formed by crystal settling in a basaltic magma, the relative amounts of olivine and plagioclase in the solid phase differed markedly.

If the lavas were derived from a crystalline differentiate, re-fusion of the magma must have been complete, for no relic feldspars were noted in the sections examined. The even grain size of the Minto plugs, which cooled at comparatively shallow depths in the volcanic conduits, is evidence of complete liquidity of the magma. The remarkably uniform variation of the more andesitic lavas, particularly in Al_2O_3, would also suggest crystallization from an entirely fused melt.[32] It seems to the writer that the porphyritic nature of the High Cascade lavas is due to partial crystallization of magma from a completely liquid state under near-surface conditions shortly before extrusion, rather than transportation of relic phenocrysts from great depth in a large magma chamber.

No systematic vertical variation in composition of the High Cascade lavas is apparent, except possibly within the Minto series, or in the largest peaks in the Olallie series. Certainly no gradual change from basalt to rhyolite, or vice versa, is evident in the series as a whole, for the youngest intracanyon lavas of the entire Cascade range, including the Columbia River gorge, are as basic as the lower Outerson flows. Some may hold that the andesitic peaks of Olallie age represent final acidic differentiates, as they probably do, but it is extremely doubtful whether the bulk of these few peaks is as great as the sum of the almost

[32] N. L. Bowen: *Evolution of the igneous rocks* (1928) p. 113-124. Princeton University Press.

innumerable smaller basaltic cones, such as Larch Mountain and Olallie Butte, and the intracanyon flows of similar age. The youngest, postglacial flows and cones of the Cascades are characteristically dark gray rather highly olivinitic basalts.

The upper part of the Minto series contains a number of pyroxene andesite flows, and at least one hornblendic flow. Whether the hornblendic plugs and lavas southwest of Olallie Butte belong in the Minto or Olallie series is not evident; in either case they were erupted shortly before or soon after the end of Minto time. These acidic lavas suggest that differentiation had progressed during Minto time, but are by no means conclusive evidence, as andesites also occur in the Outerson series.

Although the youngest High Cascade lavas in the region north of Mt. Jefferson are not consistently more acid than the older flows, they do show much more variety, and in this way indicate progressive differentiation. They do not show the extreme range of composition found in the Lassen Peak region,[33] yet range from rather basic basalt to dacite. The great preponderance of basalts which are chemically not far removed from plateau basalt magma and the absence of rhyolites indicate that either differentiation did not progress far, or the most acidic fractions were not erupted. The eruption of rhyolites in the Lassen Peak and Crater Lake regions suggests that the former alternative is the more probable. This seeming lack of differentiation in the northern part of the Cascades as compared with the south end has been noted before.[34] The evidence of the Minto plugs indicates that the magma was completely fluid at comparatively shallow depths. If this is true, similar conditions probably obtained at moderate and, possibly, at great depth, and differentiation by crystal settling could not proceed. Presumably a crystalline differentiate very rich in iron, with moderate amounts of magnesia and lime, and a small amount of alumina (olivine, augite and magnetite ?) was separated early in the magmatic history to form the magma of which most of the flows were formed. The occurrence of large andesitic cones literally in the midst of small basaltic cones suggests that the more acidic lavas were erupted from comparatively small pockets or cupolas in the main magma chamber. The localization of differentiation may have been due to local cooling, shown by the porphyritic nature of the andesites and associated basalts in the high peaks as compared to the uniformly fine-grained textures in the basalts of the smaller cones. The evidence that great quantities of basaltic lavas were erupted while the large andesitic peaks were being built is indisputable, whatever the mechanism may have been.

[33] Howel Williams: *Geology of the Lassen Volcanic National Park, California,* Univ. Calif. Pub. Bull. Dept. Geol. Sci., vol. 21, no. 8 (1932) p. 195-385.

[34] Arnold Hague and J. P. Iddings: op. cit., p. 222-235.

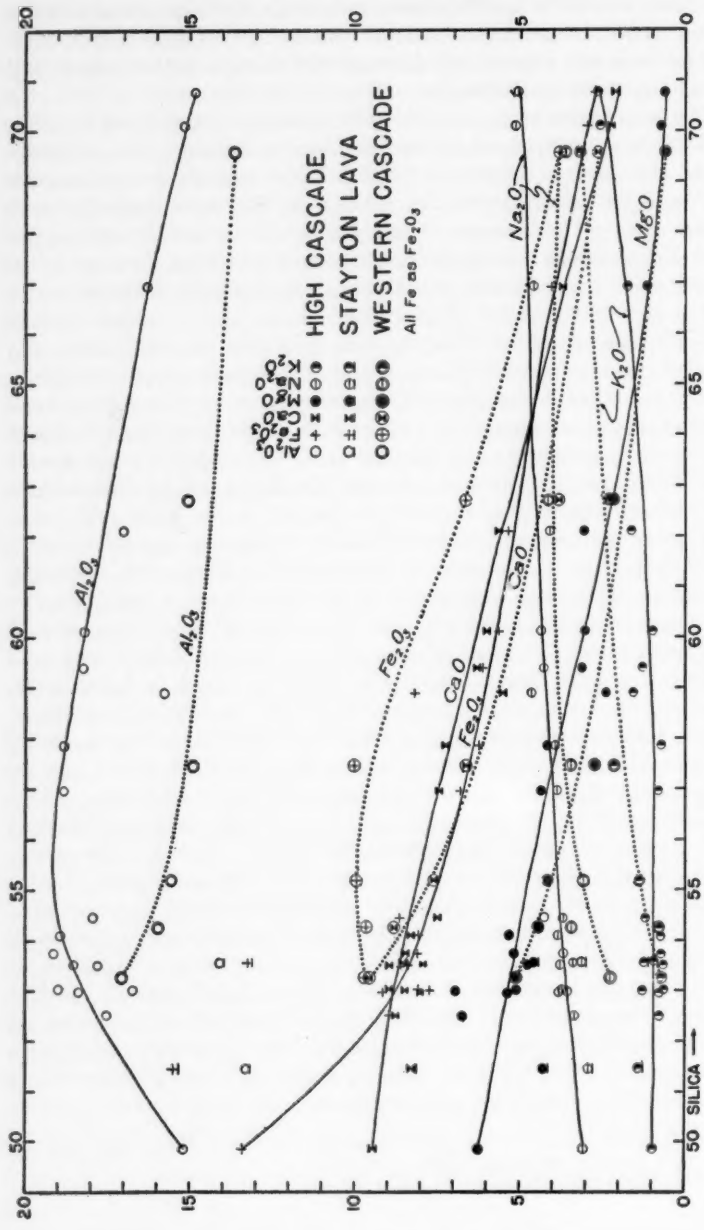


FIGURE 3.—*Composition-variation diagram of Oregon Cascade lavas*

OREGON CASCADE LAVA VARIATION DIAGRAM

A glance at a composition-variation diagram of Oregon Cascade lavas (Fig. 3) reveals striking chemical differences between the Western Cascade and the High Cascade lavas. These differences have already been briefly pointed out by Callaghan and Buddington.[35] The younger lavas are richer in Al_2O_3 in the range above 53.5 per cent SiO_2, in lime above 55 per cent SiO_2, and in NaO_2 above 65 per cent and below 56 per cent SiO_2. The Western Cascade lavas are consistently richer in iron oxide and potash. There seems to be little, if any, consistent variation in magnesia between the two series.

The smoothed curves indicate a break in the magmatic history at the end (?) of Miocene time, or at the same stratigraphic position as the structural break at the top of the Western Cascade sequence. The break may not be so sharp as it appears from the available data. It seems significant that the three analyses of low alumina High Cascade basalts, excluding the more basic Cupola Rock lava, are all of Outerson, or basal High Cascade, lavas. The andesite plotted at 58.9 per cent SiO_2 is likewise an Outerson flow. These Outerson lavas, whose only systematic variation from the High Cascade curves seems to be in alumina, appear to be intermediate between the Western Cascade and the true High Cascade lines of descent. Possibly their inconsistencies are a result of mixing of partially differentiated magma by the diastrophic movements at the end of Western Cascade time. How diastrophism could effect the indicated changes in the course of magmatic differentiation through High Cascade time is a question on which little information is available.

It is noteworthy that the Outerson series contains a much greater proportion of fragmentals than any other group in this part of the High Cascades, implying greater evolution of volatiles during eruptions. It is conceivable that stirring or mixing of a magma, with the resultant disturbance of equilibria, might contribute to the escape of gases in unusually large quantities during subsequent eruptions.

The Stayton basalts fit neither High Cascade nor Western Cascade composition curves, particularly with regard to alumina and iron. Their unusually high iron content and relatively low alumina—iron actually exceeds alumina in the black basalt—suggest that they were derived from plateau basalt magma by separation of plagioclase, probably with magnesian olivine, as the black basalt is also low in magnesia. The crystal fraction in early stages was apparently very different from that in the normal Western Cascade sequence. The lavas show no evidence of being hybrid types, but appear normal in every way. Their compo-

[35] A. F. Buddington and Eugene Callaghan: *Dioritic intrusive rocks and contact metamorphism in the Cascade Range in Oregon*, Am. Jour. Sci., 5th ser., vol. 31 (1936) p. 441.

sition suggests that differentiation during the Miocene followed two divergent courses in contiguous areas; plagioclase was the dominant crystalline component of the Stayton magma while magnetite and olivine were crystallizing from the normal Western Cascade or Sardine magma. Differentiation of the Sardine magma, incidentally, progressed to the point where large quantities of rhyolites were produced. It would appear that the course of differentiation in the Cascade magmatic province is not nearly so simple as the available information has indicated heretofore.

HIGH CASCADE TYPES OF VOLCANICS IN WESTERN NEVADA

In the course of mapping of the Comstock Lode District by members of the United States Geological Survey, a series of lavas with strong High Cascade affinities was discovered in the Virginia Range, immediately to the east and southeast of Reno, Nevada, on the Carson topographic sheet.

The volcanics vary greatly in thickness as they were extravasated on a mature surface; at most they are a few hundred feet thick. They include fragmental rocks only in very subordinate amounts. The lavas are best exposed and exhibit most variation in the area east of Lousetown Creek, about 6 miles north of Virginia City; the name "Lousetown series" has accordingly been proposed for them. In the type locality the lavas are characterized by modified dip slopes displaced by normal faults. Similar lavas form conspicuous dip slopes on the western side of the southern end of the Virginia Range, northeast of Carson City. Small remnant patches of these lavas occur at numerous localities in the mountain range, and a similar lava caps a part of the Flowery Range in the Tallapoosa district, about 15 miles northeast of Virginia City.

The Lousetown volcanics lie on other volcanics of upper Miocene or lower Pliocene age. A major unconformity between the two series is indicated by the fact that the older rocks are gently folded and in certain areas are highly sheared, bleached, and cut by small veins, whereas the younger lavas are fresh and are apparently broken only by the latest faulting. It is suggested that the Lousetown lavas may be similar in age to the Minto series.

The lavas of the Lousetown series in the type area are all olivine basalts. Fresh pyroxene andesite in the southern portion of the Comstock Lode District quadrangle has been tentatively correlated with the basalts on the basis of analogous structural relations and petrologic similarity.

The basalts range from pale- or medium-gray porous lavas to fine-grained glassy dark flows. All types have a characteristic sheen on cer-

tain surfaces in hand specimen, owing to fluidal arrangement of the feldspar laths. Minute olivine grains or pseudomorphs are usually visible megascopically, although in one olivine-bearing specimen none could be detected even with a good lens. Prominent pyroxene phenocrysts are rather common in the coarser varieties.

In thin section the basalts are very similar. Between 0.5 and 2 per cent of every slide examined consisted of olivine or of pseudomorphs after olivine. All have good flow structure, the black glassy types showing it best. Phenocrysts of labradorite and clinopyroxene are common. The groundmass is usually intergranular and contains granular or prismatic hypersthene and augite, tabular calcic plagioclase, and granular magnetite in a colorless glass base. Dendritic magnetite in the glass gives the darker lavas their color.

The gray basalts were probably erupted from a plug or plugs of the granular Minto type. Such a plug is well exposed in the fault scarp which forms the east slope of hill 6540, east of Lousetown Creek and about 6½ miles north northeast of Virginia City. The exposed portion of the plug is not more than 150 yards across, and extends only two-thirds of the way to the top of the hill.

The corerock is a rather pale-gray olivine micro-gabbro. It differs from the typical Minto micro-norite in that it contains 1 or 2 per cent of olivine and very little hypersthene. The plagioclase has the same extreme zonation from labradorite to oligoclase or albite, and forms about 75 per cent of the slide. Cristobalite is also present. For the most part the micro-gabbro is somewhat porous but not truly vesicular. In places, however, the rock is extremely vesicular and has gas bubbles three-eighths of an inch or more in diameter. The relation of some of the vesicular zones suggests that volatile constituents escaped upward along rather definite channels, some of which were only a few feet wide.

Careful examination of lava—micro-gabbro contacts by F. C. Calkins and the writer failed to reveal any sharp boundaries. The relation was everywhere gradational, the change from plug rock to lava occurring in most places within a distance of 10 feet or less. No dikes or apophyses indicative of an ordinary intrusive relation were noted, nor were any inclusions of wall rock found in the plug. In short, this neck in the Lousetown series is identical with the Minto plugs of the High Cascades in Oregon insofar as the relation to adjacent lavas is concerned, and its petrographic nature is strikingly similar.

The Lousetown pyroxene andesite is much like the hypersthene-cristobalite andesite of the Battle Ax and the Minto series. It is more or less platy and in places develops good columnar structure at an angle of about 45 degrees with the flow banding. In hand specimen, smutty-

appearing, elongate, pyroxene-magnetite pseudomorphs are more or less conspicuous, and cristobalite is recognizable in many cavities. Large rounded hypersthenes are totally absent. In thin section the resemblance to analogous Battle Ax lava is complete. The texture is porphyritic intergranular, phenocrysts being mainly hypersthene prisms and labradorite. The groundmass consists of an abundance of prismatic hypersthene with some augite, the pyroxenes being interstitital to tabular feldspar. The augite-magnetite ghosts are in every way identical with those previously described. High-temperature silica is an important element in the groundmass.

The presence of lavas of the High Cascade type in western Nevada indicates an extension of the Cascade petrologic province farther south and east than it has hitherto been recognized. The earlier Tertiary lavas and intrusive rocks of the Comstock Lode District are likewise petrologically very closely related to those of the Western Cascades. Lassen Peak, 120 miles to the northwest, is the generally accepted southern limit of the Cascade province. The Virginia and Flowery Range occurrences suggest that other outlying areas of Cascade-type lavas may be found in unsuspected districts.

CONCLUSIONS

The lavas of the Cascade Mountains in Oregon are divided into two main groups, the Western Cascade and High Cascade, on the basis of a marked unconformity and consistent differences in chemical composition.

The Western Cascade lavas, probably Miocene and older, have been folded, faulted, intruded by dioritic plugs, and locally mineralized. Petrologically they consist of two main interfingering types, the Stayton lavas and the Sardine lavas.

The Stayton lavas are black and gray basalts similar in many respects to the Columbia River basalts. Both gray and black basalts contain more or less olivine, are characterized by ophitic textures, and have a salic : femic ratio of 1.25 : 1 to 1.4 : 1. The gray basalts contain some cristobalite.

The Sardine lavas, about 6000 feet thick, range from olivine basalt to rhyolite, but andesites predominate. The most common lava type is basaltic andesite characterized by glomeroporphyritic textures and containing as much as 75 per cent (by volume) feldspar. The salic : femic ratio, in contrast with Stayton basalts of similar silica content, is over 2 : 1.

The High Cascade lavas are divided into several groups on the basis of structure, but are petrologically homogeneous. All these lavas are very fresh and apparently retain initial dips. The sequence as a whole,

except the large pre-Wisconsin peaks such as Mt. Jefferson, is composed mostly of rather light-gray olivine basalts. Pyroxene andesites and hornblende-pyroxene andesites are represented by comparatively few flows. The basalts are characterized by light color, platy fracture and superficial resemblance to andesites. Hypersthene is usually present, and olivine is probably present in nearly all lavas containing less than 55 per cent silica; the average basalt contains 52.5 to 53.5 per cent SiO_2. The outstanding feature of the lavas is the high salic : femic ratio of 2.5 : 1 which is reflected in a maximum feldspar content of 68.5 weight per cent. The basalts were extruded from micro-norite plugs containing both olivine and cristobalite, as do some of the basalt flows. The andesites, some of which are dacitic in composition, are characterized by hypersthene and large amounts of cristobalite and tridymite.

No systematic variation of lavas from basic to acid or vice versa was noted in either Western or High Cascade sequences. In the Western Cascade lavas lateral variation is fully as great as vertical. In the High Cascades basalts predominate throughout, but andesites are probably most abundant in late Minto and Olallie time.

The High Cascade lavas are in general richer in Al_2O_3, CaO, and Na_2O and poorer in FeO and K_2O than those in the Western Cascades. The smoothed curves of composition variation diagrams indicate a rather sharp break at the end of Western Cascade time marked by changes in the curves for all elements, but particularly for alumina and iron oxides. The Stayton basalts lie on neither High Cascade nor Western Cascade composition curves, and apparently reflect complications in the course of magmatic differentiation.

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MOVEMENT OF SOUTH CRILLON GLACIER, CRILLON LAKE, ALASKA

BY BRADFORD WASHBURN AND RICHARD GOLDTHWAIT

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INTRODUCTION

During the past century a great number of ice-movement experiments have been made in various parts of the world. Some investigators have concentrated on a physical and chemical analysis of the behavior of glacier ice under stress. Others have focused their interest upon observations of glacier motion, advance and retreat over monthly, yearly, or seasonal periods. A host of others merely have made notes on the visible changes taking place from time to time in the dimensions of present-day glaciers. Many photographs, both on the ground and from the air, have been taken to record permanently these variations.

Forbes, Agassiz, and Tyndall were particularly intent upon studying the great seasonal and monthly variations of the glaciers of the Alps. Tarr, Martin, Chamberlin, Von Engeln, Hess, and many others have more recently been conducting field work bearing upon the physical nature of ice movement and on the shearing, thrusting, and apparent plastic deformation which is inseparably related to glacier flowage.

In the course of the geological field work of the Harvard-Dartmouth

Alaskan Expeditions of 1933 and 1934, during which the base was at Crillon Lake in the Fairweather Range, a series of detailed glacier-motion observations were made at the suggestion of Professor J. W. Goldthwait, of Dartmouth College. This more intensive study of the mass

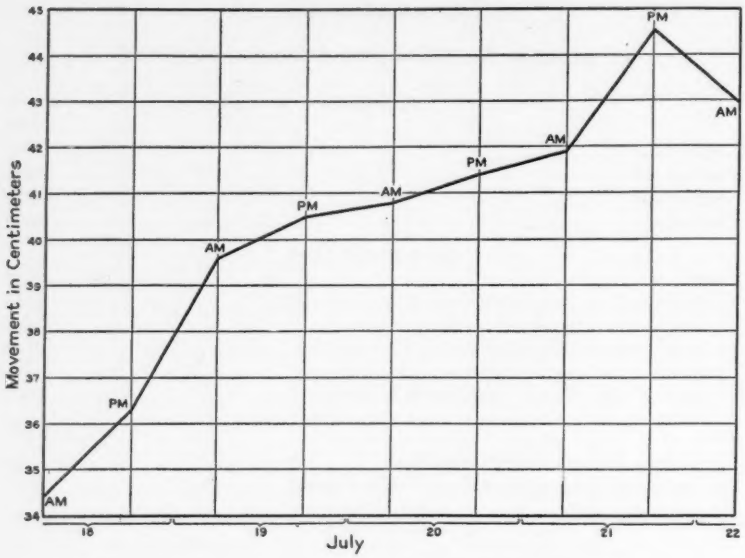

FIGURE 1.—*South Crillon Glacier, 1934 semi-daily movement*
Showing effect of unbroken cloudless weather, July 17-21.

movement of glacier ice aimed at a more complete understanding of its unknown physical properties. The earlier experiments in the analysis of ice deformation dealt with the nature and extent of the relative movements within the mass of a glacier, whereas the present observations measure the movement of the entire ice mass referred to an immovable point not on the glacier's surface.

During the first summer, R. P. Goldthwait made a series of bi-hourly observations of the movement of the South Crillon Glacier, which clearly demonstrated that its ice moved with marked fluctuations in velocity and also suggested the existence of possible diurnal periodicity in flow.

The following year, as a result of the many questions raised by this initial set of observations, new and more elaborate equipment was flown to Crillon Lake and a group five observers recorded the advance of the South Crillon Glacier, hour by hour, both day and night, for eight days.

Weather records were kept and a remarkably close correlation was found between air temperature, insolation, and the rate of ice movement.[1]

After a detailed study of the graphs and weather records compiled from the field notes of the 1934 expedition, many questions still remained unanswered, although several new and definite facts about glacier motion were discovered. These facts were, in brief, as follows:

1.) The rate of movement of a typical Alaskan valley glacier is irregular, varying from zero up to a maximum of nearly 6 centimeters an hour.

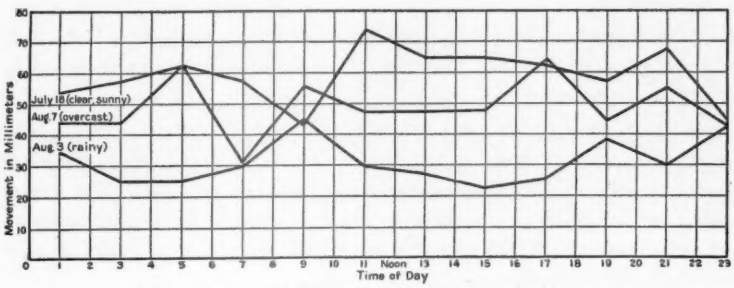

FIGURE 2.—*South Crillon Glacier, bi-hourly movement*
Record of July 18, August 3 and 7, 1934.

2.) There is an immediate and marked effect of prevailing weather on glacier motion. An ice mass flows faster during warm, sunny weather than when the sun is overcast (Figs. 1 and 2). The minimum of movement is reached during a cold rainstorm (Fig. 2).

3.) Several periods occur during the course of the day, when motion is faster than at other times (Fig. 2). These periods, whose length and intensity vary greatly from day to day, are usually three or four in number:

a) Shortly after midnight;
b) Around eight o'clock in the morning;
c) Around noontime;
d) Later in the afternoon.

4.) It is especially significant that, despite its far-reaching effect on the total daily movement of a glacier, the weather produces no noticeable change whatsoever in the relative speed of the ice, hour by hour

[1] For a detailed discussion of the glaciological research of these two expeditions, *see* Bradford Washburn and R. P. Goldthwait: *The Harvard-Dartmouth Alaskan expeditions of 1933 and 1934*, Royal Geogr. Soc., Geogr. Jour., June (1936).

or at different periods of the day. Weather variations result merely in a general increase or decrease of total movement, the graph pattern for a rainy day being almost identical with that for clear weather.

Although these four facts, gleaned from a study of the 1934 hourly graphs, were a definite advance in the knowledge of glacier behavior, it was still quite impossible to tell whether the movement recorded by any given observation had taken place:

a) Slowly and steadily throughout the hour, or

b) In the form of several marked jerks or crescendos of speed during the hour, or

c) In the form of myriad, irregular, tiny thrusts or jerks which, scattered through the hour, resulted in the hour's advance as recorded by the instrument.[2]

These observations had shown, beyond doubt, that the movement of a glacier is peculiarly irregular and possibly even spasmodic. They had failed, however, to demonstrate conclusively the nature of this irregularity or periodicity. In order to reach a more definite understanding of ice movement in minutiae, it appeared necessary to make another series of observations.

It appeared after a careful examination of the record of the 1934 expedition that accurate observations on the same glacier could easily be made at 30-minute intervals. If an especially high-powered instrument could be obtained and a new marker devised to put on the glacier, there was an excellent possibility that, in clear weather, 15-minute observations would not overrun the limits of instrumental accuracy.

During the spring of 1936 the Geological Society of America set aside a grant from the Penrose Bequest, partly to defray the expenses of a party of three, which was to make a new series of observations at Crillon Lake during June and July of that year. The three men chosen to carry out this field assignment were Russell Dow, of Woodsville, New Hampshire; Benjamin R. Twiss, of Englewood, New Jersey; and David I. Brink, of Minneapolis, Minnesota. Dow had previously played an active part in the Crillon Glacier observations of 1934.

PREPARATION

The light "Tavistock" theodolite reading to seconds of arc, used by the 1934 expedition, was replaced by a Wild precise geodetic theodolite (Pl. 1, fig. 1), loaned by Hamilton Rice, of the Institute of Geographical Exploration at Harvard. The entire success or failure of the 1936

[2] In 1934 a Tavistock theodolite, manufactured by Cooke, Troughton, and Simms, York, England, reading with no verniers direct to seconds of arc, was used.

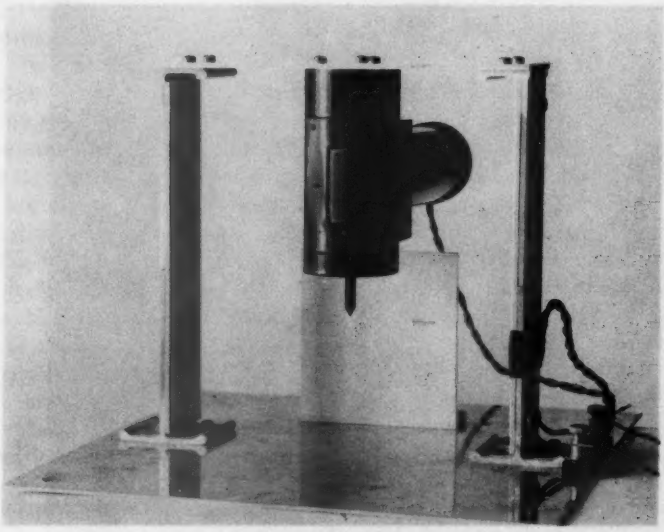

FIGURE 1. GLACIER-MOTION MARKER

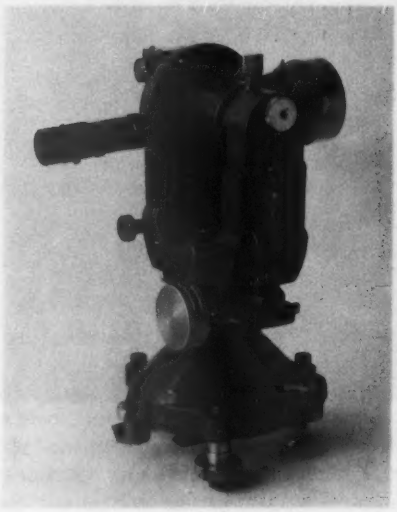

FIGURE 2. WILD PRECISE THEODOLITE

INSTRUMENTS USED ON 1936 EXPEDITION

AERIAL VIEW OF TERMINUS OF SOUTH CRILLON GLACIER

Small circles mark location of instrument and glacier-motion marker. Width of glacier, approximately 2000 feet; depth of ice at point of observations, approximately 350 feet; Distance from observation station to marker, 1000 feet.

observations hinged on the care and operation of this amazing instrument. Its telescope has a magnification of forty times, with a remarkably clear and brilliant image. Its horizontal and vertical circles, graduated on glass, read directly, with no interpolation or verniers, to tenths of a second of arc. A unique optical system permits one to read both circles through the same micrometer eyepiece by means of a simple prismatic mirror which may be changed instantly from one to the other.

In order to benefit from the high degree of precision made possible with this theodolite, a new target or motion marker (Pl. 1, fig. 2), especially designed for both day and night observations, was manufactured by the Mann Instrument Company of Cambridge. This marker consisted of a brass tube (painted black) 4¾ inches long and 2¼ inches in diameter, suspended rigidly in a vertical position from a small bridge of aluminum angle. This bridge was firmly screwed to an aluminum base plate three thirty-seconds of an inch thick and 12 inches square. The vertical tube, which was carefully sealed at both top and bottom, had a small vertical slit-opening, 1½ inches high and three-sixteenths of an inch wide, covered by a frosted celluloid window, to be observed at night and a quarter-inch brass rod, one inch long and tapered to a point projected from the bottom of the tube, for daytime observations. Another section of similar tubing was soldered to the marker tube in the form of a "T" projecting backward directly behind the frosted window. A 4-volt, .75-ampere bulb was mounted at the end of this tube and a small lens focused its rays brilliantly onto the window. All joints were soldered or thoroughly shellacked to prevent the heavy Alaskan rains from penetrating this lighting unit. The whole instrument was made as simply and as ruggedly as possible.

One of the greatest difficulties and dangers of the earlier observations—the morning and evening trips across the glacier to detach and to connect the lighting circuit—was eliminated in 1936 by laying 1000 feet of double-strand, rubber-coated wire, supplied by the Simplex Wire Company, of Cambridge, which linked the marker directly to the observing station where contact was made on a small knife-switch to light the marker before each night-time observation. The resistance imposed by this amount of wire necessitated the use of seven dry cells at the observing station in order to light the bulb in the marker. The bulb was lighted to only about two-thirds its rated brilliance in order to reduce halation about the aperture in the marker and to make observations more accurate.

A series of field tests, carried out in Cambridge during May, proved this assembly capable of consistently and correctly detecting movements as small as two millimeters at a distance of 1000 feet. Assumption of

any greater degree of accuracy than this in glacier-motion experiments is absurd because of the slight slumpings of the marker which must necessarily take place on account of ablation.

OBSERVATIONS

On July 19, 1936, the three observers flew to Crillon Lake from Port Althorp, Alaska, and re-established the former base-camp near the end of the glacier, on its eastern shore. The next five days were spent in making camp and in the installation of the apparatus in the field.

The smoothest point on the surface of the glacier was chosen for the marker, which was screwed to the bottom of a heavy, white box inverted on the ice. A recording thermograph was placed beside the box on the glacier. The theodolite was fitted upon a special aluminum and brass base permanently mounted in the head of an 8-inch column of concrete poured directly on a rough bedrock surface. A tarpaulin was rigged above this assembly to protect it from both sun and rain. The wire from the marker was led up the gully and a touch from the small switch illuminated the instrument for night observations. Another wiring system within the theodolite, controlled by a rheostat at its base, illuminated both its graticule and the circles for readings after dark. A tent, 7 feet square, with a wooden floor, was set up a few yards from this station on a steep slope (Pl. 2), and an alarm-clock, constantly set in advance, kept the observers on their job.

On June 25 at 9:00 a.m. the series of half-hourly observations began. The glacier's movement was recorded and graphed in the field in seconds of arc as observed on the theodolite. A simple calculation gave the linear advance of the ice (Fig. 3). This series of observations continued unbroken, with the three observers working in shifts, until 8 o'clock on the morning of July 14. From the conclusion of this 30-minute series until 3:45 p.m. on July 16 a series of readings was made at 15-minute intervals with successful results (Fig. 4). Throughout the entire period of twenty days, eleven observations were missed, several occasioned by oversleeping, and several by the slumping of the marker in the warm sun, requiring a speedy trip out onto the glacier to readjust it firmly, and the resumption of observations after an hour. There was no break in the series of quarter-hourly observations.

Although clear weather prevailed during most of this period, a moderate rainfall on June 30 and heavy showers on July 6, 7, and 9 provided the variety hoped for as a medium for the study of weather conditions on the movement of the ice. Unfortunately, however, this bad weather was neither severe enough nor of long enough duration to produce as marked an effect on the ice motion as did the heavy rains of August 1934.

CONCLUSIONS

It is not the intention of the writers to attempt to make any all-embracing statements in explanation of the various phenomena suggested by these limited studies of ice flow. In order to reach a fuller understanding of the effect of external conditions, such as insolation, temperature, and precipitation, on glacier motion it is believed that it will be necessary to carry out an extended series of bi-hourly or even semi-daily observations in other localities over a period of at least six weeks to two months. Such a program would not only prove expensive but would tax the ability and patience of the observers. It would probably not, however, turn out to be any greater strain than the observers experienced at Crillon Lake in 1936, reading a micrometer at half-hour intervals for nearly three weeks.

The surveys of 1934 and 1936, however, have resulted in the discovery of several phenomena with regard to the mass flow of ice, which should be of value to anyone contemplating further investigation in this field of research. These phenomena, together with a few remarks and general observations on the results of this field work, follow:

1.) The movement of the surface of a glacier is extremely jerky. This fact, first suggested by Chamberlin's surficial experiments in the Alps and Alaska[3] and later by the hourly records of 1934, seems definitely proved by the half-hourly and quarter-hourly observations of 1936 (Figs. 3 and 4). This irregularity reached its peak with a movement of 20 mm. for a single quarter hour on July 15, 1936, 7:00 to 7:15 p.m. (Fig. 4), although the average is approximately 6 mm. per fifteen minutes. At another time, 8:00 to 8:15 a.m. on July 15, 1936, the total fifteen-minute motion was only a single millimeter (less than the accepted limit of instrumental accuracy). On certain other days not graphed in this paper, absolutely no movement at all was recorded during several fifteen-minute periods.

Either the errors in observation of such tiny movements from such a distance are distorting a continuous, gently fluctuating rate of glacier flow, or the interval between the quarter-hourly observations is approximately equal to the short period of rest between sudden spasms of glacier motion. The increasing irregularity of movement as shown on graphs based on shorter and shorter intervals of observation suggests that small instantaneous spasms of motion affect each portion of the ice surface at least once every fifteen or twenty minutes.

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[3] R. T. Chamberlin: *Instrumental work on the nature of glacier motion*, Jour. Geol., vol. 36 (1928) p. 1-30.

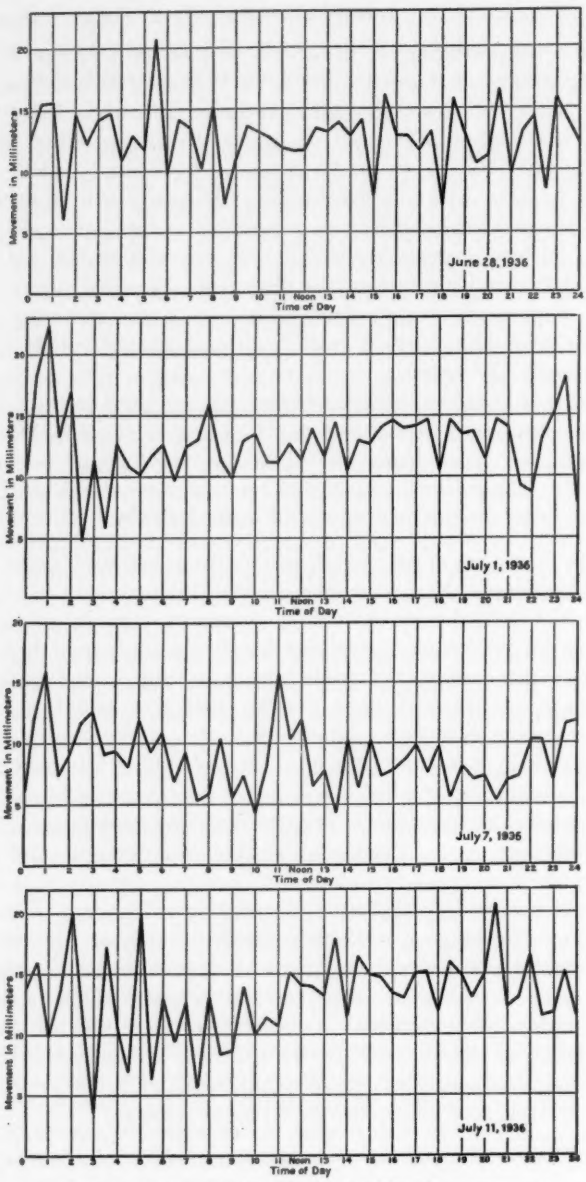

FIGURE 3.—*South Crillon Glacier, half-hourly movement*
Record of June 28, July 1, 7, and 11, 1936.

2.) Large variations in the rate of motion are notably less during the afternoon hours than during the evening and early morning. The total movement from 11:00 a.m. to 3:00 p.m. averages one-third to one-half that from 11:00 p.m. to 3:00 a.m. The comparatively smooth and uni-

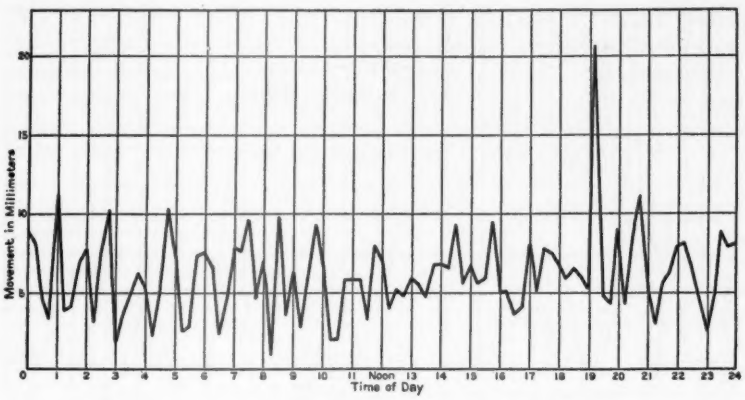

FIGURE 4.—*South Crillon Glacier, quarter-hourly movement*
Record of July 15, 1936.

form night and afternoon movements are most commonly followed by jerky, impulsive motion after dawn and sunset. This instability seems to be closely correlated with the daily effect of the sun's heating.

3.) A curious and completely unexplained periodic effect is seen by bi-hourly graphs of the 1934 and the 1936 observations. At least three, and usually four, periods of especially rapid motion were recorded each day. These were normally of approximately two hours' duration, but the hours when they occurred varied slightly from day to day. In general, all bi-hourly graphs seem to show periods of greater movement shortly after midnight, about 7:00 a.m., and 2:00 p.m. and again early in the evening. The two night periods and that which came between 6:00 and 8:00 a.m. occurred consistently; the early afternoon advance sometimes was at noon or slightly earlier, and sometimes was entirely absent. If absent, however, the early evening acceleration was likely to be more marked and to take place slightly earlier than usual. Effects of these variations can be seen on the graphs of the half-hourly movements (Fig. 3).

4.) The great drop-off in motion during night hours (6:00 p.m. to 6 a.m.) is perhaps the most regular and consistent phenomenon to be

seen throughout the motion graphs of each year as compared with the total of daytime motion (6:00 a.m. to 6:00 p.m.) (Fig. 5). Eighteen of the twenty-four available semi-daily records show the daytime movement to be greater by the ratio of approximately 4 to 1. Not one of

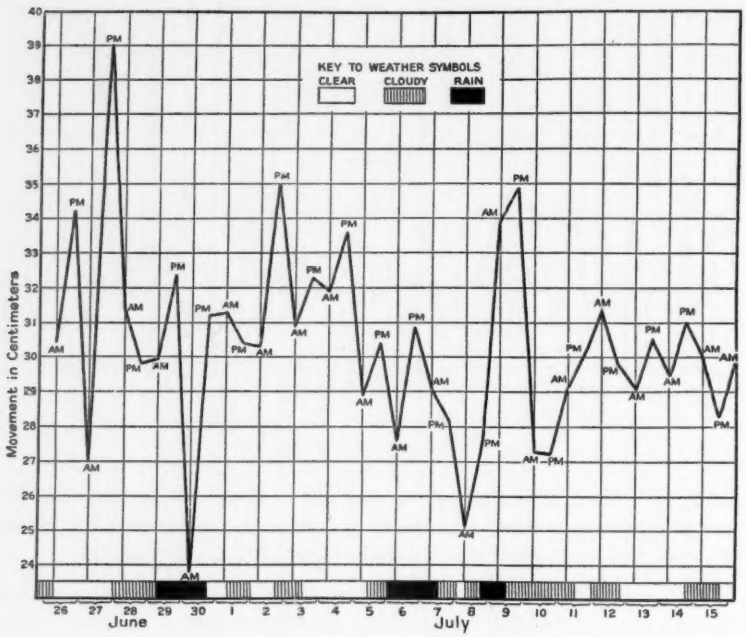

FIGURE 5.—*South Crillon Glacier, 1936 semi-daily movement*
Record of June 26 to July 15.

the six exceptions occurred on the two days when the average temperature for the night happened to be higher than that for the preceding or following day; all, however, occurred during cloudy weather. The greatest differences between day and night motion occurred on sunny or on partly cloudy days. Toward the end of each rainy period or immediately following it, a single day of unusually small movement is recorded. Possibly the five cases on the records are not sufficient to substantiate this phenomenon as a fact, but a definite correlation between prevailing weather and glacier motion is suggested.

5.) Definite response, however, is shown to rainy or sunny periods by the remarkably consistent acceleration of glacier motion from July 18

to 21, 1934, during continuous fair weather (Fig. 1). Over this period of four days an increase in daily movement from 70 to 86 cm. is found, with a sudden drop at its end, when cloudy weather resumed. Conversely, the average daily motion decreased from 65 to 60 cm. during the intermittent cloudy weather from June 26 to July 15, 1936 (Fig. 5). It is admittedly recognized that not every short rainy period during these observations brought an attendant marked or prolonged decrease in glacier motion. This may be explained by the fact that none of the rainstorms encountered during the 1936 observations seemed to last long enough to result in as great a general slowing down of the ice as is seen in the rainy period surrounding August 3 to 4, 1934 (Fig. 2).

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LOWER ALGONQUIN BEACHES OF CAPE RICH, GEORGIAN BAY

BY GEORGE M. STANLEY

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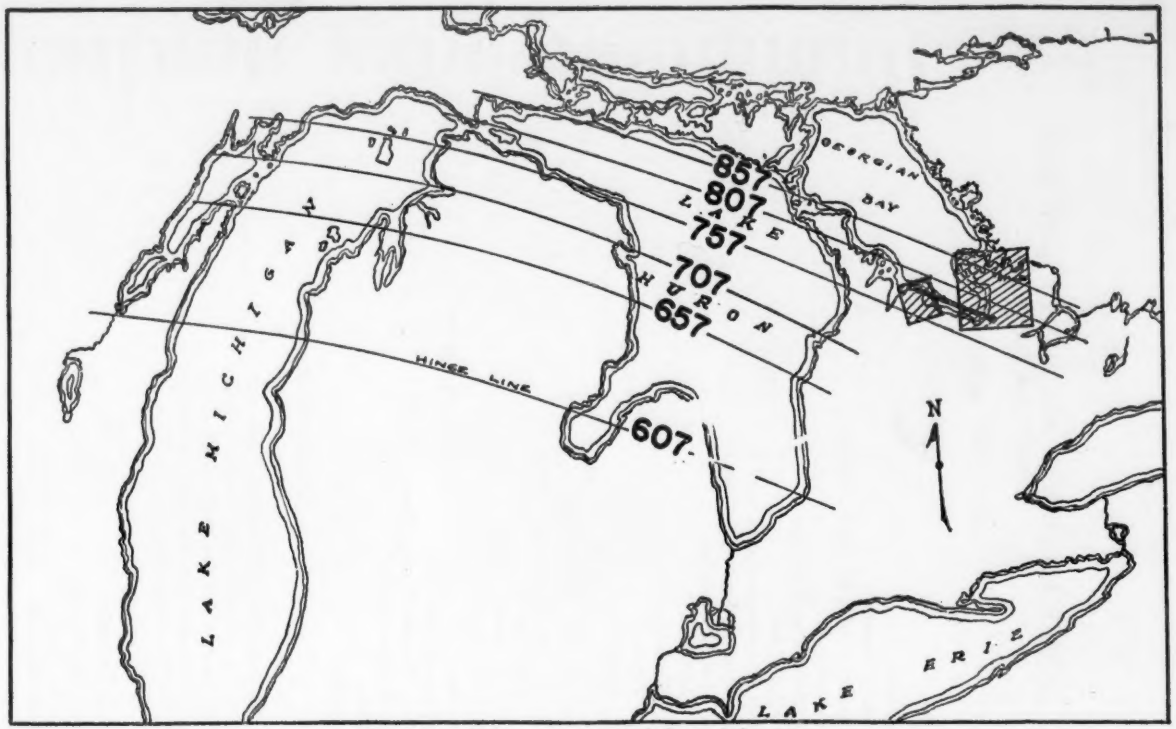

FIGURE 1.—*Index map of central Great Lakes area*

Hachured rectangles give location of study areas; the smaller represents area of Figure 2, studied in 1936; the larger is the area of 1935 studies around Penetanguishene. Double line between rectangles connects Cape Rich and Wasaga Beach localities where the Payette shore is impounded by the Nipissing. Curved isobases represent the deformed Algonquin water-plane with elevations above sea level, as adapted from Goldthwait.

INTRODUCTION

Discovery of essential parallelism among the Algonquin and lower Algonquin water-planes of Georgian Bay necessitates considerable revision in later Great Lakes history. The lakes would seem to have drained northeastward to the St. Lawrence–Ottawa sea during the several lower Algonquin stages, and long before the Nipissing channel (North Bay-Mattawa) was used.[1] The total number of the stages and the positions of the corresponding outlets are not known, but the lower Algonquin beaches are numerous and cover a wide range of elevation in the north. Investigation of these far-reaching beaches, about Georgian Bay, northern Lake Huron and Lake Superior, combines naturally with the tracing north and west of the four lower Algonquin water-planes already established in southeastern Georgian Bay (Fig. 1). The Cape Rich studies furnish one step toward the solution of this extensive problem.

Field work was done mostly in September, 1936, with the aid of a grant from the Penrose Bequest of the Geological Society of America, and with good assistance in the leveling furnished by Mr. Richard Bilbie, of Ann Arbor. The measurements at Vail's Point, however, were made in 1934 under a grant from the Faculty Research Fund of the University of Michigan.

The writer acknowledges the courtesy of maps from the Ontario Department of Public Highways and of local information from Mr. Robert Patterson and Mr. Walter Carson, both of Cape Rich.

PRELIMINARY DISCUSSION

PAUCITY OF LOWER ALGONQUIN FEATURES

The strength and continuity which marked out the Algonquin and Nipissing shorelines for earlier investigation by Taylor and others were the product of erosion during lake advance and culmination in level. The lower Algonquin stages (viz., Wyebridge, Penetang, Cedar Point, and Payette) seem to have had only brief activity as compared to the Algonquin or the Nipissing. Moreover, they were mere temporary halts in a declining lake and lacked the effectiveness of rising water. Registration at the lower Algonquin levels was therefore relatively weak, and fewer distinguishable features were formed.

Tracing of these beaches is confronted by a further handicap. Throughout long stretches of coast line, Nipissing wave erosion cut deeply enough to wipe out such marks as may have been produced at the immediately higher stages. This is especially so on exposed coasts which lack the protection of bed rock. The Nipissing bluff is commonly from 15 to 40

[1] G. M. Stanley: *Lower Algonquin beaches of Penetanguishene Peninsula*, Geol. Soc. Am., Bull., vol. 47 (1936) p. 1933-1960.

feet high in the 17 miles of coast from Cape Rich to Thornbury. In this distance no Cedar Point shore feature remains and the Penetang shoreline is only fragmentary. Continuous tracing of any single lower Algonquin beach for more than a mile or so is totally impossible.

THE TERRAIN

Rocky slopes and ledges of the limestone escarpment are to be seen in the high parts of the peninsula northwest of Meaford, but near the lake shore glacial drift is abundant and generally masks the Paleozoic beds. Even here the rock layers are close to, or occasionally at the surface, though not very evident. The tall bluff which marks the Algonquin shoreline, for miles to the southeast of and to the north of Meaford, appears generally at an angle of repose as though cut into glacial drift. And, indeed, much drift must have been swept away, as the many residual boulders show. In places, however, soft Paleozoic beds may be seen in the bluff, even though generally slumped down, grassed over, or otherwise obscured. The Nipissing bluff near Cape Rich, likewise, reveals an occasional exposure of the bedrock series, from which the shingle deposits of the Nipissing and post-Nipissing beaches at that locality were derived.

METHODS OF MEASUREMENT

Levels were carried over flat stretches by means of a telescopic level on tripod and a collapsible stadia rod. For rapid ascents where long sights were not possible, an ordinary pocket level was employed in the same fashion. The surface of Lake Huron–Georgian Bay (578.20 feet for 1936 season) was the starting datum for all leveling at considerably below the mean level (580.8) of the lake (Pl. 3, fig. 2). This procedure is sufficiently precise for the purpose since its errors, accruing to perhaps one foot or so, are within the limits of measurement of the shoreline features. For instance, the lines of levels for the numerous measurements at Cape Rich were repeatedly closed in a network and this was tied to the lake at three separated points, the resulting discrepancies being 0.30, 0.75, and 0.05 of a foot. Closure of several traverses of a mile or so within the net yielded generally mild discrepancies (0.20, 0.85, 0.00, 0.10, and 0.00 of a foot). Certain corrections were made for the largest, but they were generally dismissed as insignificant. From such checking is derived confidence in single lines of levels at other points and, needless to say, confidence in the entire procedure of measurement. On nearly all the abandoned shoreline features, multiple observations were made in order to render a typical average elevation. Many of the values were then taken to the nearest even foot. They represent the crests of depositional ridges, the highest wave-formed parts of the eroded terraces.

DESCRIPTION OF DATA

TABULAR SUMMARY OF DATA

Measurements are shown in Table 1 by localities, starting at Vail's Point and working east and south along the coast to Thornbury. The placing of the various beaches into columns according to lake stages is determined by their position on the profile of the tilted water-planes (Fig. 5). Elaboration of the Cape Rich data is made later. All measurements are by spirit level unless otherwise noted, and are given in feet above sea level, A.T.

DETAILS OF BEACHES AT CAPE RICH

General statement.—In view of the limited shore activity at the lower Algonquin stages, together with the widespread Nipissing destruction previously mentioned, the full development and preservation of beaches at Cape Rich furnishes the best spot in the area for detailed study. Since they display certain relations which could not be explained in the foregoing abbreviated table, further description follows, commencing with the oldest.

Algonquin.—This highest old shoreline is marked by a great bluff, a salient feature of the landscape all around this part of Georgian Bay (Pl. 1). Near Cape Rich this bluff stands 200 feet or more above the Algonquin terrace. It lies 1½ miles or so west of Cape Rich. The highest certain Algonquin shoreline is a gravelly bar (801) at the southeast corner of Mountain Lake; a slightly higher flat (804), across the lake, is not as surely a beach. Lower ridges, down to about 760, are found nearby, and these, together with the highest beach, constitute what Taylor[2] has called the upper group of Algonquin beaches. A mile or more of the Algonquin shoreline to the east of Vail's Point has been removed by heavy erosion at the present level (Fig. 2 and Pl. 1, fig. 2) which is an unusual circumstance in this region.

Wyebridge.—A great hairpin-shaped bar lying one mile west of Cape Rich marks the next water-level below the upper Algonquin beaches. Four observations on the south limb average 717.5, the greatest discrepancy being 0.5. Nine observations on the north limb, average 718.0 and lie between limits, 716.8-718.8. The apex of the hairpin projects eastward and is fringed along the east and north sides by a series of light ridges (elevations in Table 1) down to 13 feet below the main bar (718-705). The high ground on which the bar lies was later cut away to a terrace and bluff at the next stage. The position and relations of the Wyebridge, as well as later beaches, are shown in Figure 3.

[2] Frank Leverett and F. B. Taylor: *The Pleistocene of Indiana and Michigan and the history of the Great Lakes*, U. S. Geol. Surv., Mon. 53 (1915), p. 415-416.

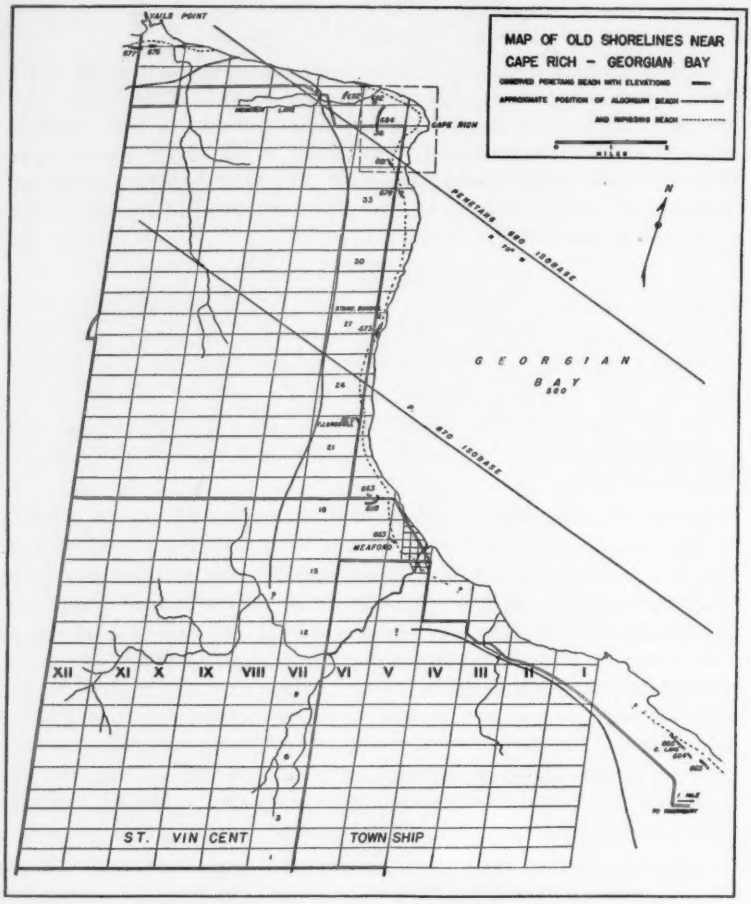

FIGURE 2.—*Map of portion of Grey County, Ontario*

Showing probably all the identifiable existing fragments of the Penetang shoreline. Isobases show at right angles the direction of tilting. The dashed rectangle about Cape Rich is the area shown in Figure 3. The Algonquin beach west and southwest of Meaford is plotted amiss on the diagram. The course is as follows: west from southeast corner of Lot 13, Concession VII, curving toward north to northwest corner of Lot 18, Concession VIII, thence toward north, three quarters of a mile west of its plotted position.

Penetang.—This shoreline is marked by a bold shingle and gravel bar half a mile west of Cape Rich road corners. The bar trends southward across the road from the bluff and terrace just referred to and is illustrated in Plate 2. Thirteen observations were taken on the crest of the

bar, between limits of 683.2-685.0, and average 683.9. The terrace at the foot of this bluff, with an elevation of 673 or 674, stands 10 or 11 feet below the bar, at about the same level as the distinct little bench (676) which was cut into its front slope. Another Penetang beach (681) lies a mile farther south, and two less conspicuous forms (both 682) are just to the north.

Cedar Point.—About 600 feet east of the Penetang bar just described, and trending across the road, nearly parallel to it, is another bold gravel ridge, the Cedar Point beach (Pl. 3, fig. 1). Six observations on it, north of the road, average 662.9, and lie between limits of 662.4-663.8. Abstraction of the material for this beach left no prominent bluff like that adjoining the Penetang bar. However, a bench (656) was cut into this bar also, some 7 feet below its crest (663), probably during a low-water period. At about the same level, another gravel ridge (657) was extended farther to the northeast. Near the end of this beach are two much fainter ridges at lower levels (654, 648). All these beaches are abruptly truncated at this point by the bluff and terrace of the Nipissing shore, and it is obvious that a marked change in delineation of the coast occurred between the Cedar Point and the Nipissing stages of the lake.

The impounded Payette beach.—Although least conspicuous, this beach is the most interesting one of the series.[3] Where crossing the east-west road, a quarter of a mile east of the Cedar Point beach, the Payette is a scarcely discernible ridge and rises only about a foot above the surrounding clay flat, to an elevation of 624. Its surface is not gravelly like the other beaches but is mantled with a fine clayey material in which stones are sparse. Woodchucks have brought from beneath the surface, however, sufficient quantities of nicely rounded shingle to show the true material of the beach (Pl. 4, fig. 2). In one place the beach is parallelled for a short distance in back by a secondary ridge at practically the same level. About 500 feet south of the east-west road, the principal beach merges with the southward-rising surface of the flat, and its ridged character is lost. However, it grows stronger to the north, standing fully 6 feet above the flat near the north end, where it terminates abruptly. The great Nipissing bar truncates it nearly at right angles, stands 10 feet higher, and obviously overlies it. At the junction, the marly clay surface of the Payette beach, is receiving a slow contribution of shingle talus from the steep back slope of the Nipissing bar.

[3] William Logan [*Report on the geology of Canada*, Geol. Surv. Canada, Rept. (1863) p. 911] mentions the beaches at Cape Rich. The writer's Payette beach was overlooked, presumably because of its subdued form, although it must have been visible when the other beaches were observed. Logan gives a few aneroid elevations which run into error but presents a nice, short description of the Nipissing bar which apparently evoked considerable interest.

Measurements were run continuously along the course of the Payette beach, nearly parallel to the tilt-line (Fig. 4). They show that the beach is higher at the northeast end by an amount that agrees with the tilt rate (about 2.75 feet per mile), but the difference in this short half-mile extent of the beach is not necessarily due to tilt.

The position of the Payette beach, behind the Nipissing and at a lower level, as may be seen in the plan and section of Figure 3, indicate plainly enough that it is older than the Nipissing, and was in all probability formed at a lower lake stage. Its clay surface must have been added to the shingle nucleus of the beach in quiet waters, protected by the Nipissing bar to the east. The generally low, inconspicuous form of the Payette ridge is probably a result of two things. It was washed down mildly by the Nipissing waters, and also the fine debris from these waters perhaps had some tendency to smother the ridge, which may have stood up more boldly on the surrounding flat before this blanket was added. No good sections of the beach were exposed for critical inspection.

The feature may well be called an "impounded beach"—that is, a beach submerged and protected behind a higher beach that has been formed at a later stage of the lake. Some special term for this seems advisable, inasmuch as it is the second such discovery in the near region and illustrates a significant relation of the water-planes. Accordingly the term retroactively applies to the marl-covered Payette gravel ridge found near Wasaga Beach, during the summer of 1935, which, in situation and history, is identical with the one just outlined.

The great Nipissing bar.—Throughout the 20 or more miles of coast between Vail's Point and Thornbury, the Nipissing shoreline is an almost continuous erosional terrace and bluff. The only principal break in this is the great bar at Cape Rich. This bar swings about in a great bulbous curve, a mile and a half in length, which connects at either end to the usual Nipissing bluff. It is less than a quarter of a mile from the water, and almost parallel to the present shore (Fig. 3). In cross section, the bar is unsymmetrical. Whereas the lakeward side slopes gently off to Georgian Bay in a succession of wave-formed ridges, the landward edge drops precipitously to the broad flat behind. The amount of fall reaches a maximum of 23 feet toward the north end of the bar but varies along its course (Pl. 5, fig. 1).

The bar reaches greatest elevation at the ends (638, 640) and declines toward the middle portion (629). This condition is graphically shown in Figure 4, and no further figures need be given here since they will be discussed later. The greater height near the terminal points implies contribution of material swept along shore from the adjacent bluffs, and the size of the shingles near the south end of the bar have the same

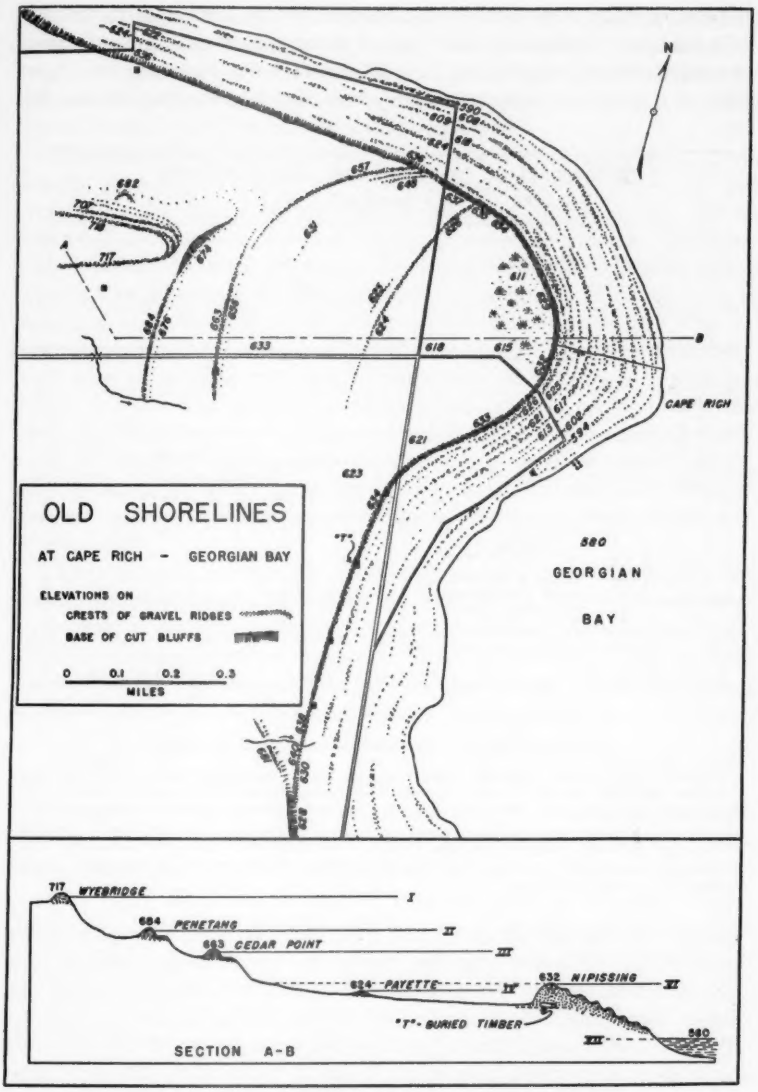

FIGURE 3.—*Sketch map of beaches in Cape Rich area*

Showing shorelines of all stages except the upper Algonquin group, which lies farther to the west. The section A-B illustrates the abrupt back slope of the Nipissing bar and, by Roman numerals, the sequence in which the beaches were formed. The fifth stage is probably too low to be shown on the section, and the buried timber is projected from its true position shown in the upper diagram. For general location, *see* Figures 1 and 2.

indication, when compared to the smaller pebbles farther north (Pl. 6).

In addition to this longshore drift of debris, there was certainly some movement directly on shore. It is known that the Nipissing lake went through a period of rising level which culminated in the Port Huron dis-

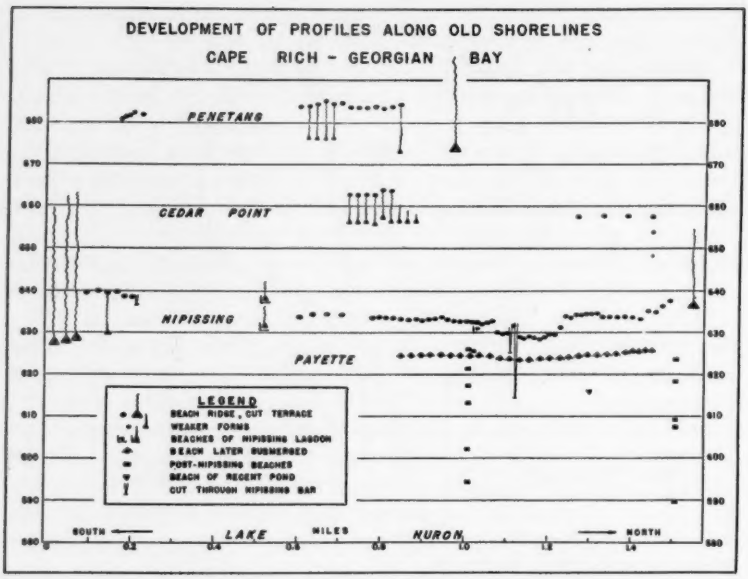

FIGURE 4.—*Altitude observations on Cape Rich beaches*

Each dot on the diagram indicates a single reading, the corresponding numerical value not being presented. Each row of dots follows the course of that particular shoreline in Figure 3 and therefore progresses in a changing and different direction from the other lines of symbols. A slight northward increase in altitude may be noticed in the Penetang and Payette beaches. Note the range in level indicated for several stages, a depositional ridge for high water, generally a bench or terrace for low water. Of the cuts through the Nipissing bar, the southernmost is a shallow creek bed, the others artificial trenches for drainage of a marsh.

charge. During this period the surf must have continually swept material up the front slope and over the crest of the bar so as to impart a rolling movement to the entire feature and drive it backward toward the land. The migration thus effected was not unlike that of a barchan. To this process is attributed the imposing development of the Nipissing as a great deep-water barrier, not only at Cape Rich, but in various places about the Great Lakes, such as Taylor describes on northwestern Mackinac Island and elsewhere.[4] The writer has witnessed this prom-

[4] *Op. cit.*, p. 453.

inence of the Nipissing shore ridge at innumerable spots, notably at Rainbow Cove on Isle Royale, near Presque Isle on Owen Sound, and elsewhere. Taylor's excellent description of the Whittlesey beach,[5] likewise the product of a rising level, can be applied almost equally well to many characteristic bars along the Nipissing shore line.

The inferred shoreward progression of the bar at Cape Rich, finds confirmation in the record of Robert Patterson's well. The well lies exactly on the summit of the ridge as indicated by the symbol "T" in Figure 3, and excavated shingles litter the surface roundabout. Although the elevation of the bar at this point was not measured, it may be safely assumed to be near 636. Mr. Patterson states that a solid timber was struck at a depth of 25 or 26 feet when the well was drilled. Whether this log drifted to its position before the Nipissing gravels were piled back over it, or whether it grew in place together with other vegetation, which was drowned by the Nipissing waters and then invaded by beach deposits, there is no way at present of telling. In the latter case, it would seem to call for a rise in water-level at this point of at least 25 or 30 feet, since the elevation of the log is about 611. Unfortunately, there is not yet any section exposed which might disclose a whole forest bed at this level.

The impounded condition and preservation of the Payette beach still further confirms the deduction that the Nipissing bar was built backward and up the shore slope. Its erection must have been a gradual and continuous accompaniment of the rise of the lake, in order to have successfully defended the Payette beach from destruction by the advancing surf.

The lagoon of Nipissing time.—The great bar enclosed a lagoon which was over a mile long and shaped much like an inverted comma. However, the lagoon was not entirely isolated from the lake until probably the closing stage of the Nipissing when the water had begun to withdraw. Assuming the former lake level along the present 634- or 635-contour at Cape Rich, nearly half of the bar, as it now stands, would have been submerged. And near the apex of its broad curve lay an open channel with a depth of 5 feet or more, through which storm waves could easily have entered the lagoon and, with considerably diminished strength, washed its landward shore. A sandy strip (631) about 0.3 mile northwest of the road corner seems to mark the shore of this lagoon, but there is more distinct indication about half a mile southwest of the road corner where two little terraces are strewn with boulders and clearly waterwashed. One of these comes at essentially the same level (632) as the

[5] *Op. cit.*, p. 382.

sandy beach just referred to (Pl. 5, Fig. 2). The other stands at 638 (four observations between 638.0 and 638.4) and must have been made when the water was higher. Indeed, it is within 2 feet of the highest Nipissing observation (640) in the vicinity, at the bar's south end, where the shingle might have been thrown up from a water-level about 6 feet lower. When this elevation is compared with that of other Nipissing shore features at Cape Rich and in nearby Georgian Bay, it seems like an extraordinarily high Nipissing water-level, and perhaps only a very temporary one. It would appear likely that upon the return of discharge to Port Huron, there was a rather rapid but small drop in level due to clearing out of forest, shrubbery, and possible sand dunes, that may have lined the channel floor, unused since Algonquin times.

The depth of wave agitation within the lagoon was necessarily less than that outside. The great bar broke the full force of the waves and allowed only a shallow disturbance to penetrate across the lagoon. With the Nipissing Lake at 634 or even 638, the Payette beach (624) lay generally below this disturbed zone and was only slightly modified, if at all.

The coating of fine debris which covers this ridge, is spread widely over the entire lagoon flat. Several gullies which enter the back shore of the lagoon seem to have brought considerable material from the upland, and the finer clay would naturally settle evenly over the lagoon floor. Waves may also have brought in fine suspended matter when the water was roiled by storms. The south part of the lagoon seems graded up with sandier sediment in a wedge-shaped alluvial fan. Its apex coincides with the southern point of the lagoon where two gullies debouch, at an elevation (640) close to that of the bar itself. The fan descends and broadens to the north. In order to escape eastward, water from one of the gullies has cut a little trench, 2 feet down into the bar, this being accomplished perhaps after completion of the fan had begun to retard its drainage northward. It may be noted that the buried log (611) lies considerably lower than the general level of the lagoon flat nearby (623). This suggests that the lagoon floor may have been built up several feet since the time when the log was first covered with gravel.

After the lake had fallen below the lowest sag in the Nipissing bar (629), the lagoon became totally isolated and thenceforth its only natural drainage was through the pores of the rubble bar. A small marsh northeast of the road corners may be considered a survivor of the lagoon. Its water surface was 611 at the time of the writer's visit, but has been higher recently for a distinct band of flotsam was found up to a point about 5 feet higher. The local inhabitants say that it frequently used to flood in spring and after rains, being a nuisance to the road and fields. This provoked years ago a curious attempt to drain the marsh by tun-

FIGURE 1. ALGONQUIN TERRACE AND BLUFF
Southwest view between Cape Rich and Meaford. The bluff stands about 200 feet above the former lake shore and exposes soft Paleozoic rocks near the top.

FIGURE 2. HEAVY EROSION AT THE PRESENT SHORE
East view near "the clay banks," probably a unique instance of Georgian Bay cutting back so deeply as to remove the Algonquin beach. The bluffs rise nearly 400 feet higher than the lake, 150 feet above the obliterated Algonquin level. The great boulder-paved terrace indicates the amount of glacial drift thus removed.

WAVE-CUT BLUFFS

FIGURE 1. SOUTHWEST VIEW OF THE GRAVEL BAR
The bar is the highest strip of open land in the distance, at elevation 684. The little bench (676) where the haystack stands was cut into its front slope during low water.

FIGURE 2. NORTHWEST VIEW OF TERRACE AND BLUFF
The bluff seems to have been the source of material for the bar in Figure 1. Toward the left, the terrace (674) merges with the bench shown in Figure 1.

PENETANG SHORELINE

nelling through the Nipissing bar, but the project failed. Two open ditches were since dug through. One was abandoned, but the other was carried to a depth of about 18 feet. Its threshold was measured at 614.5 by the writer, and apparently correlates with a distinct little bench at 615-616 cut into the shingly back slope of the bar along the borders of the marsh. Apparently this successfully drains off the ponded water so that the level no longer exceeds 615 or 616.

Other beaches.—A number of the shore forms cannot be assigned to any single stage. The upper Algonquin group of beaches has already been mentioned. This series is found below the highest Algonquin in an almost continuous succession and indicates a gradual withdrawal of the lake. Near Cape Rich it covers an interval of about 40 feet (800-762), though perhaps greater farther north. The beaches of this group were possibly due to uplift in the north while Lake Algonquin continued to discharge at Port Huron, and after outflow at Kirkfield had entirely ceased.

Similarly the succession of beaches between the Nipissing and the present lake were formed while the North Bay outlet was being elevated to its present height, and while the lake drained at Port Huron. It is the writer's impression that the post-Nipissing beaches in the Georgian Bay region occur indiscriminately at all altitudes between the Nipissing level and the present.

A number of irregular bouldery hummocks and ridges were found about a mile southwest of Cape Rich, occurring a great distance out on the offshore bottom from the Algonquin bluff. From appearances, one would naturally attribute these to some sort of lake activity. From elevations (720, 730) some of these seem too high to be associated with the Wyebridge stage, and yet abnormally low for correlation with any of the upper group of Algonquin beaches. Not enough attention was paid to these indefinite features to interpret them further.

INTERPRETATION AND DISCUSSION

PLOTTING OF RESULTS

Maps of Grey County and St. Vincent Township, prepared by the Ontario Department of Highways, were used as base for Figure 2. The isobases in Figure 1 are adapted from Goldthwait. (See bibliography at the end of this paper.) The positions of shorelines in Figures 2 and 3 are sketched in only from observation in the field, but are generally well placed by speedometer mileages where crossing roads.

The Penetang water-plane in Figure 5 is drawn according to plotted measurements, and this furnishes the only direct evidence in this area of the tilt of any lower Algonquin beaches. It agrees very well with

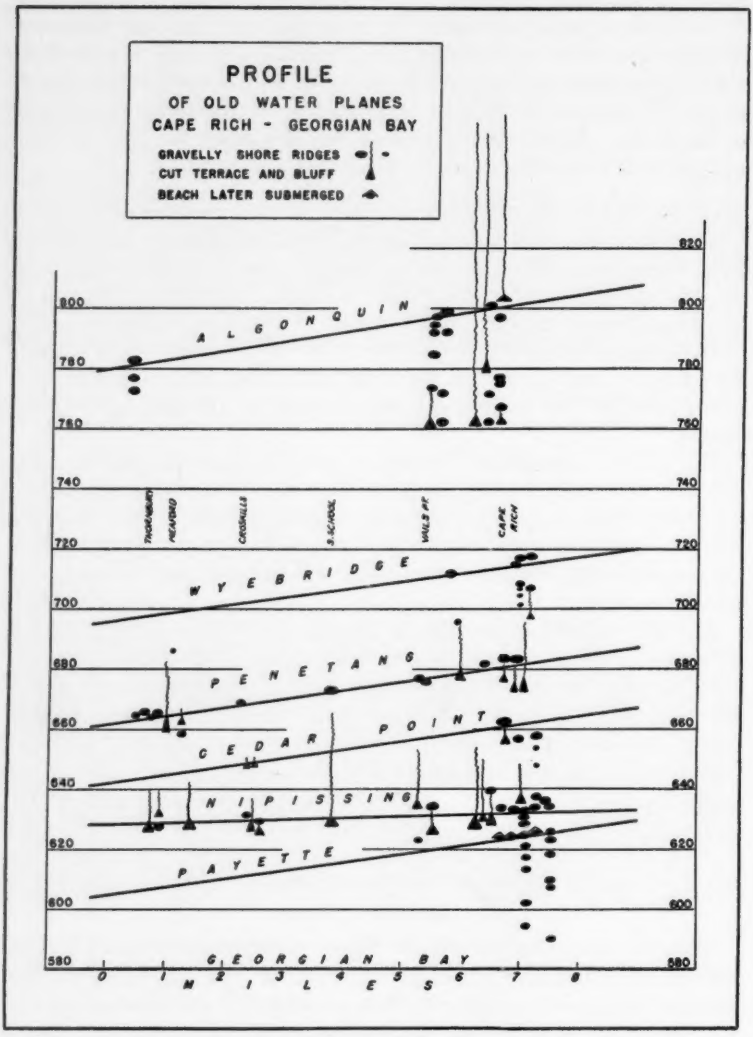

FIGURE 5.—*Profile of water-planes*

Algonquin, Penetang, and Nipissing planes are drawn to plotted measurements; the others are drawn simply in near parallel fashion as around Penetanguishene. Significant figure at center of each symbol; dots, beach ridges; triangles, erosional terraces, with a sinuous line indicating height of bluff.

the tilt of 3 feet per mile obtained near Penetanguishene.[6] The Algonquin and Nipissing planes are also drawn according to plotted measurements, and they accord with results elsewhere. The Wyebridge, Cedar Point, and Payette planes are drawn to agree with previous work by the writer and are based on Cape Rich data only in their spacing. The correlation of these water-planes with those near Penetanguishene rests on the agreement shown in Table 2.

RELATION OF NIPISSING TO LOWER ALGONQUIN BEACHES

The rise in the Nipissing water-level, which enabled the lake of that time to cut deep terraces with high bluffs, and to construct imposing gravel embankments, naturally brought about a marked adjustment or straightening of the Nipissing shorelines. By contrast, each of the lower Algonquin shorelines seems to have been feebly made during a brief halt of a declining water-level. One might say that the latter accepts the shore outline, which has been given by the pre-existing contour and terrain, whereas the Nipissing more nearly imposed its own shore outline upon the existing terrain.

As may be seen in Figure 3, the Penetang, Cedar Point, and Payette shorelines show general parallelism in ground plan. Each, as the lake fell, established a distinct beach which followed the general contour of the land and therefore trended more or less parallel to the earlier shore. The Wyebridge bar may be omitted from the comparison, since in this locality it happens to be built upon an unusually long and narrow morainic swell. A pointed headland had appeared, and in the succeeding stages it grew toward the northeast. The Nipissing lake changed this point to a broad and rounded headland or cape. During the retreat of the waters since the Nipissing, the form of the cape has scarcely varied.

Both the Cedar Point and the Payette beaches are crossed, or cut away by the Nipissing in an utter disregard for contour, for the Nipissing lies fully 20 feet below the one and 10 feet above the other. Although unscathed at this locality, the Penetang is sliced by the Nipissing in similar fashion at Vail's Point, Meaford, and Thornbury. This relation of lower Algonquin beaches to Nipissing has been sufficiently observed about Georgian Bay to be termed characteristic. The Nipissing is unconformable[7] with the earlier beaches, and there is a marked "break" between them. Exactly the same relation of unconformity, between the Nipissing on one hand, and the Penetang, Cedar Point, and Payette on

[6] G. M. Stanley: *Lower Algonquin beaches of Penetanguishene Peninsula*, Geol. Soc. Am., Bull., vol. 97 (1936) p. 1951.

[7] The writer, unaware, has used almost the very words of Taylor [Leverett and Taylor: *op. cit.*, p. 449-450] as he applied them to the relation between the Nipissing and the highest Algonquin beach where found close together in the area of horizontality.

the other, is displayed near Wasaga Beach at the head of Nottawasaga Bay.[8]

Up to this point, the "break", has been discussed as a "plan" relation of the beaches. When the water-planes are seen in profile (Fig. 5),[9] fully as striking a break appears in the divergence between the flattish Nipissing, and the steeply tilted, almost parallel, lower Algonquin water-planes. This discordance has a greater implication. Unless earth tilting was comparatively rapid during this transition, one might well suspect that it marks the lapse of a long period of time. The uplift which took place greatly exceeds the post-Nipissing deformation and likewise such movement as seems to have spread out the upper group of Algonquin beaches. Several thousand years may have been required.

The impounded beaches which mark the submergence of the Payette plane at both Cape Rich and Wasaga furnish excellent physiographic evidence of the intersection or overlapping of one lower Algonquin water-plane by the Nipissing. Perhaps more impounded beaches will be discovered where other shorelines pass beneath the Nipissing. The Cedar Point and Penetang might be expected to do so somewhere along the Lake Huron shore, and similar evidence of this may be found, if favored by topographic conditions.

EVENTS AT CAPE RICH

The succession of the various stages at Cape Rich is enumerated herewith, and is similar to the history as described for Wasaga Beach.[10]

(1) The highest Algonquin (801) beach was formed at a culmination in lake level when uplift of the Trent Valley outlet had turned the discharge to Port Huron. The lake had previously been at least 30 feet lower in this vicinity.[11] Continued uplift raised the Algonquin beach here above the Port Huron outlet, and as this proceeded gradually, a series of lower beaches were formed, down to more than 40 feet below the Algonquin. Then the lake fell abruptly about 45 feet, probably because ice gave way somewhere along the highlands east of Georgian Bay so as to open a lower passage.

(2) The Wyebridge beach (718) was now formed while the lake maintained its level for a brief time. Being weak like the three succeeding beaches, when compared to the highest Algonquin or the Nipissing, the time required to form it was relatively short. The writer might advance a guess of perhaps 200 years. Some slow erosion or change seems to have taken place, probably in the outlet, causing the lake to fall 12 feet or

[8] G. M. Stanley: *op. cit.*, p. 1950, fig. 3.

[9] Note also: *op. cit.*, p. 1949.

[10] *Op. cit.*, p. 1951-1953.

[11] Impounded beaches have been found behind the Algonquin bar at Sucker Creek, and 30 to 35 feet lower, but they will be described in a later paper.

FIGURE 1. SOUTHEAST VIEW OF CEDAR POINT GRAVEL BAR
The bar stands out against the trees beyond. Both man and horses are in the hollow, about 6 feet below, and in back of, the bar (663).

FIGURE 2. ABANDONED BOATHOUSE
Craft used to be run into this boathouse, eight years ago, but low water, ice push, and longshore drift have effectively destroyed its use. East side of Cape Rich.

VIEWS AT CAPE RICH

FIGURE 1. MATERIAL IN PENETANG BAR
The lower gravel is a bottom deposit on the off-shore slope of the Penetang bar. The upper 4 feet of coarse shingle, presumably added as a beach deposit during the closing "low water" phase of the Penetang stage, makes the surface of the bench (676) cut into the Penetang bar (683). View northeast in gravel pit 300 feet south of road.

FIGURE 2. CLOSE-UP OF THE PAYETTE BEACH MATERIAL
Only a few pebbles litter the clayey surface and one might perhaps hesitate to accept the light topographic swell as an important beach, were it not that the abundance of excavated shingles around woodchuck holes is quite convincing.

BEACH MATERIALS

more and build lower beaches. Probably ice gave way again to bring about another abrupt drop in level of about 34 feet.

(3) The next stage of the lake formed the Penetang beach (684) by remaining again at a fairly constant level. Then another quick drop of 20 feet or more occurred probably as a result of further shift in outlet.

(4) The Cedar Point beach (663) formed the third of the lower Algonquin stages. Very much like the Penetang stage, it was a relatively constant level, but involved either some fluctuation in the lake or else experienced a slight gradual decline as time went on. It appears to have endured slightly longer than the Penetang, for in general its beaches are a little stronger and more numerous. Failure of an ice dam or opening of a lower outlet introduced then another lowering, amounting to about 32 feet.

(5) The Payette beach (624) marks the last of the lower Algonquin stages to be registered at Cape Rich. From its well-rounded shingle here, and especially from its strength on Penetang peninsula, it appears to have required about as much time for formation as the Cedar Point.

At the close of the Payette stage, the lake must have dropped again, probably for the same causes as before. There is no record at Cape Rich of how far it dropped this time. There may have been several later and lower stages before the Mattawa channel was in full use. It has already been suggested that considerable time passed before the next beach at Cape Rich was formed. This is a puzzle concerning which more must be learned and the solution lies only farther north. At any rate the water reached its lowest horizon at the North Bay—Mattawa outlet, whatever its level then was at Cape Rich, and this order of events was ended.

Judging from the very slight convergence of the water-planes of these beaches as displayed on Penetang peninsula, tilting had proceeded ever so slightly during these last three or four stages. Certainly the great uplift which deformed the Algonquin and lower Algonquin beaches, followed sometime after abandonment of the Payette beach and before the formation of the Nipissing. Whether, in the main, it preceded or succeeded discharge at North Bay is an interesting speculation. In the latter case, the lake level of the time would have been surprisingly lowered.

(6) Uplift of the Mattawa Valley progressively raised the lake level on the shores to the south and west. Meanwhile a bar had been established at Cape Rich and it was built up, step by step, with the rising water. Ultimately the water topped the pass at Port Huron, and culmination was reached at Cape Rich. The Nipissing shoreline is generally considered to have been formed at this time, although many recognized Nipissing shorelines were not abandoned until sometime later.

(7) Continued uplift now gradually raised the Cape Rich area out of

the water, and a series of beaches developed along the fore slope of the Nipissing bar, at different stages, as the lake withdrew.

CAPE RICH AND WASAGA BEACH COMPARED

Although the shorelines in these two localities vary somewhat—for one lies on an exposed cape, the other at the head of a large bay—they are identical in broader meaning. The novel interpretation applied to the Payette beach and the construction of the Nipissing bar at Wasaga is paralleled at Cape Rich in precise detail. The latter locality repre-

TABLE 2.—*Vertical Spacing of Beaches Compared* *

Water-plane	WASAGA BEACH				CAPE RICH			
	I Measured elevation of beach	II Interval between successive beaches	III Elevation of water-plane	IV Interval between successive water-planes	V Measured elevation of beach	VI Interval between successive beaches	VII Elevation of water-plane	VIII Interval between successive water-planes
Algonquin	not high enough here to register Algonquin		794		801		798	
Wyebridge			?	118	718	83 / 34 = 117	714	84 / 34 = 118
Penetang	678		676		684		680	
		21		21		21		21
Cedar Point	657		655		663		659	
		34		35		39		36
Payette	623		620		624		623	
		−9		−10		−10		−9
Nipissing	632		630		630–640		632	
		52		50		54		52
Lake Huron	580		...		...		...	

* The figures for the water-planes at Wasaga are taken from the table on page 1951 of the BULLETIN for 1936; for water-planes at Cape Rich, from Figure 5 accompanying this paper.

sents the phenomenon a little better since the less-extensive Nipissing bar and lagoon have been entirely mapped, and no vexatious dunes hindered the measurements. On the other hand, no nicely layered marl was found overlying the Payette beach as at Wasaga, since there was no large river at this point to supply the calciferous waters.

At the time the parallelism of the lower Algonquin water-planes was first described, it was pointed out that the intervals between the planes might be considered practically as constants, and might facilitate identification of the planes at distant points.[12] This method should be especially suitable between areas where the total Algonquin uplift is of comparable

[12] *Op. cit.*, p. 1955.

amounts. Cape Rich lies almost on the same Algonquin isobase as Wasaga, possibly a mile or so farther north (Fig. 1). The intervals for the two localities are presented in Table 2.

The elevations for beaches in Table 2 cannot be compared quite as legitimately as the elevations for water-planes. Even in one locality the distance between successive beaches in many places is such that tilt may affect the difference in their elevations. Moreover, the original height of a depositional beach above water level (or below it) is variable and scarcely to be calculated, although a bold topographic exposure generally makes for a higher beach. This probably explains why the Payette beach is farther below the Cedar Point at Cape Rich than it is at Wasaga. At Cape Rich, the Penetang and Cedar Point bars were formed on a steeply sloping shore where the waves must have been strong; the Payette lies on flatter ground which would impede surf and prevent erection of the beach to as great a height. All in all, the differences in Table 2 are paltry. The correspondence of intervals between water-planes at Cape Rich and those at Wasaga (columns IV and VIII in Table 2) is remarkable and serves to bear out the interpretation of lower Algonquin beaches which has been made previously.

VARIATIONS IN LEVEL

The light benches formed in the front slope of the Penetang and Cedar Point bars, as well as the adjoining terrace and bar at lower levels, seem definitely a product of small variations in water-level. The abnormal difference in height between the Penetang terrace (674) and the bar (684) was probably not so great when the bar was first formed, but as the lake fell a little, the exposed terrace and bluff were eroded down to the lowered level while the crest of the bar remained intact. This lowering of the lake seems to be recorded also by a change in sedimentation on the front slope of the Penetang bar, causing coarser material to be spread over finer (Pl. 4, fig. 1). If there were no bar to be found here, and the terrace were used for correlation with some Penetang bar a few miles distant, confusion might likely arise from such lack of correspondence. Evidently the terrace and bluff cannot be expected to mark exactly the primary level of the stage it is taken to represent, since the first record has been destroyed by erosion and the creation of a later one. The situation bears out an idea expressed previously.[13]

These changes may have been due to such seasonal or secular fluctuations as occur in all the Great Lakes today. A permanent change through erosion of the outlet seems another likely cause, and, to some extent, both of these must have operated. Further complication in these stage varia-

[13] *Op. cit.*, p. 1936.

tions may have been involved through a slight amount of tilting that seems to have taken place while the strands were being registered, for there appears to be some convergence in their water-planes.[14] The slight uplift differential between Cape Rich and outlets to the northeast may have caused for a time a slight rise in water of a foot or two for each stage. This seems probable on theoretical grounds and it would have helped to define clearly each beach, but no direct evidence indicates it.

Considerable variation is found in the level of the Nipissing shoreline at Cape Rich. The slope of the water-plane (half a foot per mile) might be expected to render differences of about three quarters of a foot over the extent of the numerous Cape Rich observations. Actual disparities are great enough to make this factor negligible.

A difference of 8 feet exists between the Nipissing terrace (628) next the south end of the bar, and a similar feature (636) joining the northwest end (Fig. 3). A variation of fully 11 feet is found along the crest of the bar itself (Figs. 3 and 4). The northwest end grades so insensibly into the higher ground of the Cedar Point ridges that the exact contact was not surely spotted, although it lies within 100 feet west of the road, and at an elevation of perhaps 640. A portion of the original Nipissing bar here has doubtless been removed by the same Nipissing bluff-cutting which intersects the Cedar Point deposits. An observation of 638 was made on the crest of the bar just east of the road.

At the south end of the Nipissing bar, it is certain that some of the original bar deposit has been consumed by subsequent Nipissing erosion. This may be seen by an examination of Figures 3 and 4. The same water-level which made the 628 terrace could scarcely have thrown up the shingle ridge at 640. Indeed the latter is cut away in front by a sharp bench (630) which is at about the same level as the terrace and merges with it. Evidently the end of the bar (640) was formed at the highest, culminating stage of the Nipissing lake, and at that time probably joined a terrace about 5 feet lower. At a later date the terrace was eroded to a lower level (628) while the bench (630) was being cut into the end of the bar, and further deposits or gravelly ridges were thrown up more to the north, on the bar's fore slope. There is even some suggestion in the field that this sort of change was repeated several times at only slightly different levels; that is, the neutral point which sets off the zone of deposition from the eroded zone, has gradually migrated a slight distance northward along the bar, from its original position at the south end.

In the writer's opinion, the 636 terrace at the north end of the bar was formed at a previous and higher stage to the one at 628. Thus, there are several features here that might readily be taken as typical

[14] *Op. cit.*, p. 1951, table 3.

FIGURE 1. THE GREAT NIPISSING BAR
Eastward view of the back slope of the bar near the road in Lot 36, Concession VII. The lagoon hollow (left distance, where the rodman stands) is 11 feet below the summit (634).

FIGURE 2. BOULDERY LAGOON SHORELINE
From the amount of erosion here, as evidenced by residual boulders, the terrace (632) is no mere product of lagoon waters, but of waves sweeping in over the great bar, to the landward shore of the Nipissing lagoon.

FIGURE 1. LARGE SHINGLE NEAR THE SOUTH END
Note how exposure and weathering have roughened the shingle surfaces. This is the natural front slope of the beach (610), which has probably thus been bare since it was formed.

FIGURE 2. SMALLER SHINGLE OF THE CENTRAL PART
This recently exposed material has retained its smooth, rounded appearance. The view is at the abandoned trench through the northeast part of the bar (630).

SHINGLE MATERIAL OF THE NIPISSING BAR

of the Nipissing shore, and yet at different levels, with perhaps many years or even centuries intervening between the times when they were successively abandoned by the waves. Such discrepancies advise caution in any conception of correlations.

The many observations for elevation taken on the Nipissing bar are given only in graphic form (Fig. 4). Little more can be said in interpretation. The lowest point of the bar (629) may have been formed at the same time as the highest deposit (640) or sometime after; there are no data which settle this. Certainly the bulk of the bar was already in place before the highest gravels at the ends were laid down. But for thick woods and sumach snarls, more attention might have been given to the individual ridges which line its front slope.

Observations on the terraces at the base of the Algonquin bluff indicate that this shoreline, too, belonged to different stages of formation. Similar variations were found in it, and over a much greater range than for the Nipissing. The terrace is found both at 762 and at 781 at different points southeast of Mountain Lake. It varies from 787 to 771 within 800 feet eastward from the bar at Sucker Creek. It can be seen that these do not date with the highest beach which reaches 800 at both Cape Rich and Mountain Lake.

CONCLUSIONS

(1) The highest Algonquin beach and the upper group of Algonquin beaches, as well as the Nipissing beach and post-Nipissing beaches, are found at Cape Rich. Measurements and interpretation of these beaches agree perfectly with those of previous workers.

(2) In addition to these, there are four distinct lower Algonquin stages represented: Wyebridge, Penetang, Cedar Point, and Payette. These were probably due to drainage of Lake Algonquin by successively lower outlets to the northeast, as ice gave way.

(3) The Penetang water-plane in this area has the same steep slope which was found for it in the Penetanguishene area.

(4) The Payette plane is marked by an "impounded beach", one which was shut in behind a higher Nipissing bar, and thus was preserved from wave destruction during Nipissing time. This confirms in perfect detail the interpretation advanced for a similar feature south of Penetanguishene.

(5) A marked break or relation of unconformity exists between the Nipissing and the lower Algonquin beaches, this being equally pronounced, whether the latter are above or below the Nipissing.

(6) This break may signify the lapse of an amount of time not generally recognized before in the history of the Great Lakes.

(7) The characteristic cut terraces, generally measured or taken as representative for the Nipissing shoreline, show some variation in level and in time of formation, and generally post-date deposition of the highest Nipissing gravel ridges. The same applies to the Algonquin and other shorelines.

BIBLIOGRAPHY

Goldthwait, J. W.: *An instrumental survey of the shorelines of extinct lakes Algonquin and Nipissing, in southwestern Ontario,* Geol. Surv. Canada, Mem. 10 (1910) 57 pages.

Leverett, Frank, and Taylor, F. B.: *The Pleistocene of Indiana and Michigan and the history of the Great Lakes,* U. S. Geol. Surv., Mon. 53 (1915) 529 pages.

Logan, William: *Report on the geology of Canada,* Geol. Surv. Canada, Report, 1863.

Stanley, G. M.: *Lower Algonquin beaches of Penetanguishene peninsula,* Geol. Soc. Am., Bull., vol. 47 (1936) p. 1933-1960.

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STRATIGRAPHY OF THE JACKSONBURG LIMESTONE

BY RALPH L. MILLER

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INTRODUCTION

A discontinuous belt of dark-colored non-magnesian limestones of Ordovician age crops out along the Great Valley in New Jersey and eastern Pennsylvania. Isolated belts and patches are also infolded or infaulted in the crystalline areas of the Reading Prong. These limestones have been extensively utilized for the manufacture of Portland Cement and are frequently referred to as the "cement rock". In 1908, Kümmel applied the name Jacksonburg to this formation in New Jersey. The Jacksonburg belt crops out discontinuously across northern New Jersey into New York State. Similar limestones are present in Pennsylvania in a continuous belt from the New Jersey border to the Lehigh River and in discontinuous belts and outlying patches from the Lehigh River to Harrisburg, where the Chambersburg limestone is first recognized.

Detailed areal maps of the Jacksonburg in New Jersey were published by Kümmel [1] in 1900, and faunal studies were made by Weller [2] in 1903. In eastern Pennsylvania the studies of the argillaceous limestone belt have been largely in connection with the cement industry, and stratigraphic analysis has been only incidental. The present study was undertaken in an attempt to elucidate the stratigraphy of the argillaceous limestone in eastern Pennsylvania and to determine its relation to the Jacksonburg limestone of New Jersey and to the Leesport limestone farther west in Pennsylvania.

FIELD WORK

Field work on the problem was begun in the summer of 1932 and continued through parts of the summers of 1933, 1934, and 1935. Occasional week-ends have been spent in the field during 1936 and 1937. The quarries of the cement plants in the district were first visited and examined, but it soon became clear that complexity of structure and lack of good contacts in the quarries would necessitate the study of all available exposures.

[1] H. B. Kümmel: *Report on the Portland cement industry*, N. J. Geol. Surv., Ann. Rept. 1900 (1901) p. 9-101.

[2] Stuart Weller: *The Paleozoic faunas*, N. J. Geol. Surv., Rept. Paleont., vol. 3 (1903) 462 pages.

Consequently, detailed mapping was undertaken in 1933 and has been completed as far west as the type section of the Leesport formation on the Schuylkill River. Reconnaissance mapping has been carried on from the Schuylkill River to Harrisburg. Many of the important localities described by Kümmel and Weller in New Jersey have been visited.

ACKNOWLEDGMENTS

The writer wishes first to record his indebtedness to G. Marshall Kay, of Columbia University, under whose direction the work has been pursued and whose fund of published and unpublished information on Ordovician stratigraphy has been available. He is indebted to the previous workers on this problem, whose published reports contained much valuable material. Special mention must be made of the work of H. B. Kümmel and Stuart Weller in New Jersey, of Benjamin L. Miller in the Lehigh Valley of Pennsylvania, and of G. W. Stose and A. I. Jonas in the Lebanon Valley.

The writer is further indebted to managers and chemists of numerous cement companies for information about their properties. B. F. Howell very kindly lent a collection of Jacksonburg fossils, made in New Jersey by Gilbert Van Ingen. Finally, it is a pleasure to acknowledge the assistance derived from a research grant of the American Association for the Advancement of Science, conferred by the Pennsylvania Academy of Science.

HISTORICAL REVIEW

Many authors have mentioned or described briefly the Middle Ordovician limestones of New Jersey and eastern Pennsylvania, but most of these reports have been broad in scope or economic in character, with the result that little space could be devoted to the stratigraphy of the limestones. The early publications have only historical value now, for their contributions have been included in later works. Rogers in 1858 [3] and Frederick Prime in 1878 [4] and 1883 [5] recognized the presence of middle Ordovician limestones in eastern Pennsylvania and correlated them with the Trenton of New York State. H. Cook recorded similar findings for New Jersey in 1868.

In 1900, H. B. Kümmel [6] published a *Report on Portland Cement Industry of New Jersey*, accompanied by excellent outcrop maps of all

[3] H. D. Rogers: *Geology of Pennsylvania, a government survey*, vol. 1 (1858) 586 pages. Philadelphia.

[4] Frederick Prime: *On the discovery of Lower Silurian fossils in limestone associated with hydromica slates and on other points in the geology of Lehigh and Northampton counties, eastern Pa.*, Am. Jour. Sci., 3d ser., vol. 15 (1878) p. 261-269.

[5] Frederick Prime: *The geology of Lehigh and Northampton counties*, 2d Geol. Surv. Pa., Rept. Prog. D[3], vol. 1 (1883) 283 pages.

[6] H. B. Kümmel: *op. cit.*

known Trenton limestone occurrences. A year later, Kümmel and Weller [7] discussed the Trenton in their report on the *Paleozoic limestones of Kittatinny Valley*. Three years later, Stuart Weller's final report on the *Paleozoic faunas of New Jersey* [8] contained faunal lists of the best Trenton fossil localities and a detailed section of the formation at the town of Jacksonburg, Warren County, in the northwestern part of the State. In 1908, Kümmel [9] applied the name Jacksonburg to these New Jersey limestones which had been previously called Trenton in both New Jersey and Pennsylvania.

In Pennsylvania, the names Nisky and Nazareth were used by Wherry [10] and Miller [11] for lower and upper divisions of the cement formation, and Nazareth and Lehigh by Peck [12] for approximately the same lower and upper units. Later, Miller [13] referred to these divisions as the "cement limestone" and the "cement rock". Recently, it has become customary to apply the name Jacksonburg to the cement formation of Pennsylvania.

A brief summary of the published information up to this point may be advantageously inserted here. The Jacksonburg limestone of New Jersey and eastern Pennsylvania is said to be of Black River and Lower Trenton age and to rest unconformably on Beekmantown or Upper Cambrian dolomitic limestones. At some localities in New Jersey the lower beds are composed of a heavy basal conglomerate, believed locally to attain a thickness of 100 feet. In the southern occurrences in New Jersey and in the Lehigh District of Pennsylvania the lower beds are crystalline high-grade limestone and the higher beds fine-grained argillaceous limestone. The argillaceous limestone grades upward into the overlying Martinsburg shale.

In 1927, G. W. Stose and A. I. Jonas [14] discussed these limestones, particularly the areas west and south of those already considered. They were in doubt as to the equivalence of these western limestones and applied the name Leesport to the argillaceous limestones between the Lehigh and the Susquehanna rivers, with the type section at West Leesport on the

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[7] H. B. Kümmel and Stuart Weller: *Paleozoic limestones of Kittatinny Valley, N. J.*, Geol. Soc. Am., Bull., vol. 12 (1901) p. 147-164.

[8] Stuart Weller: *op. cit.*

[9] A. C. Spencer, H. B. Kümmel, J. E. Wolff, R. D. Salisbury and Charles Palache: *Description of Franklin Furnace quadrangle, N. J.*, U. S. Geol. Serv., Atlas no. 161, Franklin Furnace Folio (1908) 27 pages.

[10] E. T. Wherry: *The Early Paleozoic of the Lehigh Valley district, Pa.*, Science, n. s., vol. 30 (1909) p. 416.

[11] B. L. Miller: *The mineral pigments of Pennsylvania*, Pa. Topog. Geol. Surv., Rept. 4 (1911) 101 pages.

[12] F. B. Peck: *Preliminary report on the talc and serpentine of Northampton County and the Portland cement materials of the Lehigh district*, Pa. Topog. Geol. Surv., Rept. 5 (1911) 65 pages.

[13] B. L. Miller: *Mineral resources of the Allentown Quadrangle, Pennsylvania*, Pa. Geol. Surv., 4th ser., Topog. Geol. Atlas no. 206, Allentown Quadrangle (1925).

[14] G. W. Stose and A. I. Jonas: *Ordovician shale and associated lava in southeastern Pennsylvania*, Geol. Soc. Am., Bull., vol. 38 (1927) p. 505-536.

Schuylkill River. They recognized an unconformity at the base of the Leesport, which unconformity appears at the base of the Martinsburg where the Leesport is absent.

The writer concurs with most of the views advanced by earlier workers. Additional information, some of which was not available to previous investigators, now makes it possible to correlate more definitely with the Mohawkian standard column and to set forth the sequence of events in Mohawkian time with greater precision. In several instances, for reasons which will be discussed, the writer is in disagreement. Much of the region is structurally so complex and paleontologically so barren that it would be surprising if some conflicting interpretations had not arisen. One of the early students of the region expressed himself as follows:

> "Level as the general surface may be, it is the planed-off section of as gnarled and twisted a piece of the earth's crust as can be found in any country. Although these plications are comparatively small they are of the same nature as the gigantic overthrown anticlinals of the Alps and Apennines."[15]

REGIONAL SETTING

The sections of New Jersey and Pennsylvania in which the Jacksonburg limestone is present lie in two physiographic provinces—the Reading Prong extension of the New England Upland and the Great Valley of the Folded Appalachians. The Reading Prong consists of a mountainous belt of pre-Cambrian crystalline rocks, extending from New England southwestward to a point a few miles beyond Reading, Pennsylvania. Folded and faulted into this pre-Cambrian area are a number of belts of lower Paleozoic sediments, along which valleys have been developed. Locally the valleys separate the complex crystalline mountains into a series of quite distinct ridges, whose summits preserve remnants of the Tertiary Schooley peneplane. The Ordovician Jacksonburg limestone is associated with other lower Paleozoics in some of the valleys, but, in others, only Cambrian sediments are now preserved.

The Great Valley of the Folded Appalachians is hemmed in on the southeast by the Reading Prong and on the northwest by a belt of lower Silurian conglomerate (Shawangunk conglomerate) which forms Kittatinny Mountain in New Jersey and Blue Mountain in eastern Pennsylvania. The valley has a maximum width of 18 miles but averages about 10 miles. The Cambro-Ordovician sediments are arranged in belts along the valley, the oldest bordering upon the crystallines on the southeast, the youngest lying at the foot of Kittatinny and Blue mountains on the northwest. Locally, however, this orderly arrangement of formations is disrupted by major folding and faulting.

[15] Frederick Prime: *The geology of Lehigh and Northampton counties*, 2d Geol. Surv. Pa., Rept. Prog., D[3], vol. 1 (1883) p. 54.

A generalized section of the lower Paleozoic formations as they appear on the maps of this report follows:

Shawangunk conglomerate—Tuscarora sandstone (Silurian)
Martinsburg shales, 3000-12,000 feet
Jacksonburg limestone, 0-700 feet
Kittatinny limestones, 3500 feet
Pre-Cambrian crystallines and Hardiston quartzite

The Hardiston quartzite of Lower Cambrian age has been included with the pre-Cambrian on the geologic maps of this report, for physiographically it is more intimately associated with the mountain-forming crystallines than with the valley-forming Kittatinny limestones. The Kittatinny formation is a terrane of light-colored magnesian limestones in which faunas of Lower Cambrian, Upper Cambrian, and Beekmantown age have been recognized. In Pennsylvania, the Kittatinny limestone has been subdivided, and local names have been applied. This thick limestone series forms a youthfully dissected lowland along the southeast side of the Great Valley, whose upper surface is correlated with the Somerville erosion surface developed on the Triassic rocks of New Jersey. The Martinsburg shales, estimated in different places to be from 3,000 to 12,000 feet thick, are of high Trenton and Cincinnatian age. Along the northwest side of the Great Valley, they form a belt of rolling hills which rise above the Somerville erosion surface to accordant elevations known as the Harrisburg surface. Where the sequence is normal, the Jacksonburg (and Leesport) limestones crop out in a narrow belt at the foot of the escarpment formed by the Martinsburg shale hills. The Jacksonburg (and Leesport) is somewhat less soluble than the Kittatinny limestone so that the region it underlies tends to be at a slightly higher elevation than the Somerville level of the Kittatinny. However, the topographic break between

TABLE 1.—*Mohawkian standard column*

TRENTON	Gloucester Collingwood Cobourg Sherman Fall Hull Rockland
BLACK RIVER	Chaumont Lowville Pamelia

the Kittatinny and the Jacksonburg is indistinct or non-existent in most places, whereas the contact between the Jacksonburg limestone and the Martinsburg shales can be drawn with considerable accuracy on the basis of topography alone. Figure 1 of Plate 1 shows the physiographic expression of these three formations at one of the few localities where the Jacksonburg forms a distinct surface above the level of the Kittatinny.

The standard column of the Mohawkian, as worked out by Raymond in Ontario and New York and recently elaborated and summarized by Kay, is reproduced in Table 1. Where possible, correlations of the Jacksonburg will be made with this column.

JACKSONBURG FORMATION IN NEW JERSEY

TYPE SECTION

Near the northwest side of the Great Valley, an anticlinal belt of Cambro-Ordovician limestone forms a continuous lowland for 24 miles in northwestern New Jersey. Paulins Kill drains this lowland westward into the Delaware River. Structurally the valley is a gentle anticline, but many faults are present along the margin, as shown by Kümmel's mapping.[16] The light-colored Kittatinny limestones underlie most of the lowland. Along both margins of the valley, however, pure dark-colored limestones crop out between the Kittatinny magnesian limestones and the Martinsburg shale. The most extensive and most fossiliferous exposure of the dark limestones is at Jacksonburg (Fig. 1, no. 1), where Weller[17] described the section and fauna under the name Trenton limestone. Kümmel[18] subsequently selected this as the type locality for a new formation.

At Jacksonburg the lower beds are not exposed at present, nor is the upper contact with the Martinsburg shales. In 1903, under Weller's direction, a trench was dug which uncovered the beds continuously from the base upward for 122 feet. It was not practicable to continue the trench to the contact with the Martinsburg, estimated to be about 20 feet higher. The trench long has been filled in, but the ledges which crop out on the hillside 200 yards northwest of the mill still form the most fossiliferous exposures of the Jacksonburg in New Jersey.

Weller's description and analysis of the Jacksonburg section and fauna were thorough, so that only a brief résumé and discussion of his findings is necessary. The lower 58 feet of the rock is a nearly pure limestone formation characterized by *"Dalmanella subaequata"* and *Leperditia fabulites* (Conrad), which he correlated with the Black River of New York State. These beds also contain *Rafinesquina alternata* (Conrad),

[16] H. B. Kümmel: *op. cit.*
[17] Stuart Weller: *op. cit.*
[18] H. B. Kümmel: *Report on the Portland cement industry*, N. J. Geol. Surv., Ann. Rept. 1900 (1901) p. 9-101.

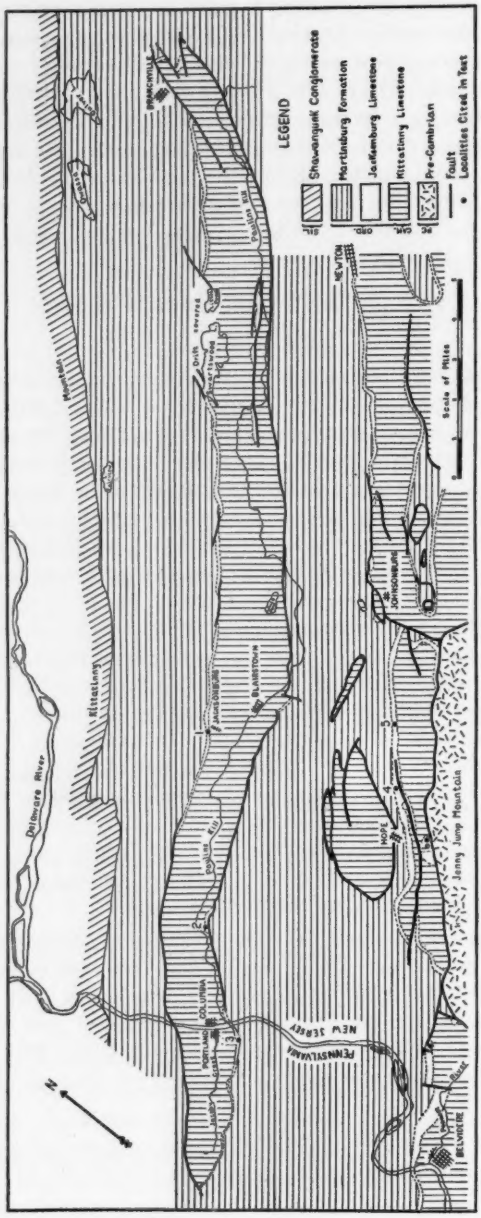

FIGURE 1.—*Geologic map of part of northwestern New Jersey*

Boundaries of the Jacksonburg in New Jersey compiled from published maps by H. B. Kümmel

"Bumastus trentonensis" and *Isotelus gigas* deKay, found also in the higher beds of Trenton age.

The Black River equivalence of these beds is questionable. *Leperditia fabulites* occurs in the Black River but is also abundant in lower Trenton limestone in the Mohawk Valley and Ontario. Weller's *"Dalmanella subaequata"* is *Doleroides* sp. cf. *D. gibbosus* Billings. Kay [19] has recently shown that beds in New York State containing *Doleroides gibbosus* (*"Dalmanella subaequata"*), which had previously been called Black River, are of Rockland age. This evidence, together with information from other localities in New Jersey, suggest that the *Leperditia* beds of the Jacksonburg section are also of Rockland age. The faunal change is so abrupt and striking between these lower 58 feet of beds and the next succeeding layers, which are referable to the Hull, that a hiatus is suggested. If the lower beds are Rockland, the hiatus would be within the lower Trenton.

Weller's "Black River" zone is followed by his "Trenton" zone which is characterized by an abundance of *Sowerbyella* sp., *Dalmanella rogata* (Sardeson), *Zygospira recurvirostris* Hall, and *Calliops* (*Pterygometopus*) *callicephalus* (Hall). The lower 28 feet 9 inches of the "Trenton" contains *Parastrophina hemiplicata* (Hall), *Streptelasma corniculum* Hall, and *Dinorthis pectinella* (Emmons). In this zone, which Weller refers to the lower Trenton, *Sowerbyella* is abundant and *Dalmanella* comparatively rare. *"Prasopora simulatrix"* Ulrich is listed as rare at a horizon 17 feet above the base. The first appearance of *Parastrophina hemiplicata* and the rarity of *Prasopora* indicate this to be correlative with the Hull of the New York-Ontario standard column (Pl. 1, fig. 1). The remaining 40 feet contains an abundant fauna in which *Dalmanella* is dominant over *Sowerbyella,* and pelecypods, gastropods, and trilobites are much more frequent. Among the diagnostic species are *Prasopora orientalis* Ulrich, which is common in beds near the top, *Cryptolithus tesselatus* (Green), and such infrequent but important trilobites as *Eoharpes ottawaensis* (Billings), *Amphilichas trentonensis* (Conrad), and *Encrinurus trentonensis* (Walcott). In addition to the 64 species listed by Weller, the writer has collected the following at Jacksonburg, all of them from the upper beds:

Carabocrinus sp.
Strophomena conradi Hall and Clark
Plectorthis (*Austinella*) *whitfieldi* (N. H. Winchell)
Whitella sp. cf. *W. subcarinata* Ulrich
Cuneamya truncatulata Ulrich
Spyroceras sp.
Holopea excelsa Ulrich and Scofield
Hormotoma gracilis Hall
Bumastus billingsi Raymond and Narraway
Dalmanites (*Achatella*) sp.

[19] G. M. Kay: *Stratigraphy of the Trenton Group,* Geol. Soc. Am., Bull., vol. 48 (1937) p. 255.

The presence of *Prasopora orientalis* in some abundance, and of *Cryptolithus tesselatus*, indicates that these beds are of basal Sherman Fall age. The top of the Jacksonburg is not exposed but the missing beds are approximately 10 feet thick.

OTHER LOCALITIES IN PAULINS KILL LOWLAND

Other exposures along the Paulins Kill Lowland are less satisfactory. One important fossil found at several other localities but not at Jacksonburg is *Receptaculites occidentalis* (Salter). At Hainesburg the *Receptaculites* beds occur in a small ledge 50 yards north of the westernmost cut of the Lackawanna Railroad (Fig. 1, no. 2). The writer has collected *Receptaculites* from beds of Hull age at Swartztown Creek near Amsterdam, New York. Kay notes that *Receptaculites* occurs in the Rockland in Ontario localities but has been found only in the Hull in the Mohawk Valley.[20] Weller stated that *Receptaculites* seemed to occur at the top of his "Black River" horizon and below the *Parastrophina* zone. If it is clearly associated with *Leperditia fabulites* and *Doleroides* (*"Dalmanella subaequata"*), this would prove that these species do not evidence a Black River age of the basal Jacksonburg. The *Receptaculites* beds of the Jacksonburg thus appear to be lowermost Hull, overlying the *Leperditia* beds of probable Rockland age.

The limestone belt, known as the Paulins Kill Lowland in New Jersey, crosses the Delaware River into Pennsylvania, where it is drained by Jacoby Creek. A quarry, 0.8 mile south-southeast of Portland, gives the only good exposure of the Jacksonburg in the Pennsylvania extension of the belt (Fig. 1, no. 3). The occurrence supplies supplementary information to that obtained at the type locality. The section measured is as follows:

	Feet	Inches	To base of formation Feet
14. Martinsburg.			
13. Covered	38	..	108
12. Gray limestone with abundant *Pachydictya* sp. cf. *P. acuta* Hall, *Rhinidictya* sp., *Strophomena* sp., and *Dalmanella rogata* (Sardeson)	13	1	70
11. Crystalline massive-bedded dark-gray limestone	5	10	57
10. Dense, fine-grained limestone	1	10	51
9. Knobbly, thin-bedded limestone containing *Isotelus gigas* deKay, *Sowerbyella* sp., *Rhynchotrema* sp., *Strophomena* sp., *Proetus* sp.; scattered conglomeratic pebbles of underlying limestone in basal bed.	3	10	49
8. Massive bed of fine-grained gray conchoidally fracturing limestone	2	3	45
7. Knobbly beds of crystalline gray limestone containing fragmentary fossils	8	11	43

[20] *Op. cit.*, p. 256, 262.

	Feet	Inches	To base of formation Feet
6. Dark-gray crystalline limestone with fragmentary fossils; *Doleroides* sp.	5	10	36
5. Knobbly, gray limestone containing *Pachydictya* sp., and crinoidals	4	..	30
4. Fine-grained gray limestone containing *Streptelasma* sp., *Doleroides* sp., gastropod indet	5	..	26
3. Crystalline gray limestone in knobbly beds 2 to 6 inches thick, containing abundant crinoid stems of all sizes, *Doleroides* sp., *Leperditia fabulites* Conrad, *Scenidium anthonensis* Sardeson, and *Monotrypa* sp.	1	6	21
2. Covered	42	..	20±
1. Beekmantown.			

The lower 27 feet of Jacksonburg exposed in this section seems to be the equivalent of the *Leperditia* beds of the type section. At the top of this zone a disconformity occurs, shown by a change in lithology and by the presence of a few rounded conglomeratic pebbles of the underlying bed. This is the hiatus inferred from the abrupt faunal change at Jacksonburg, other evidence for which was lacking at that locality. The extreme abundance of the bryozoan *Pachydictya* sp. cf. *P. acuta* Hall in the highest beds exposed in this quarry is of interest but seems to have no stratigraphic significance, as this form is also quite abundant in the beds below the disconformity.

The base of the Jacksonburg is not exposed, but loose blocks in the field show that the lowermost beds are a coarse conglomerate composed of angular blocks of the magnesian Kittatinny limestone, up to 6 inches in size, in a black limestone matrix. The toal thickness of the Jacksonburg present would appear to be 110 feet plus or minus 20 feet, but too great reliance cannot be placed on this figure, as changes of dip may occur in the covered area between the Jacksonburg quarry and the Martinsburg shale hills.

The Jacksonburg is not everywhere present along the margins of the Paulins Kill Lowland, but Kümmel's mapping indicates that its absence is explicable by faulting. Morainal deposits of the Wisconsin ice-sheet mantle much of this region so that there are not many outcrops, and long stretches exist where presence or absence of Jacksonburg cannot be established.

CONGLOMERATE IN THE JACKSONBURG FORMATION

Cambro-Ordovician limestones occur in many places in New Jersey near the margin of the pre-Cambrian belt, or folded and faulted into the crystalline area. In the limestone lowland just south of the town of Hope

(Fig. 1), a thick conglomerate is intercalated in the Jacksonburg limestones. Kümmel described this occurrence as follows:

"Beds of the conglomerate are found at a number of localities. . . . The fragments of magnesian limestone and chert range up to one foot in diameter, although the bulk of the material is only three or four inches in diameter. The fragments are well rounded and have suffered considerable transportation. . . . These conglomerate layers are apparently not at the base, but are interbedded in the Trenton limestone, the base not being shown. Exposures, however, are not sufficiently numerous to settle beyond doubt whether these beds are actually interbedded in the limestone, or whether they are really basal layers faulted into this position."[21]

A small knoll on the southeast side of the Hope-Johnsonburg highway, 3.1 miles southwest of Johnsonburg (Fig. 1, no. 5), shows 15 feet of these conglomerate beds resting apparently conformably on 20 feet of gently dipping limestone containing *Sowerbyella* and fragmentary bryozoa. At this locality there seems no possibility of a fault contact between the two, the conglomerate occurring in the Jacksonburg formation, and not at its base. It differs from the basal conglomerate observed elsewhere, in consisting of well-rounded cobbles rather than of angular blocks or fragments. Other outcrops also show the conglomerate beds apparently overlain and underlain by non-conglomeratic Jacksonburg limestone, but the exposures are not so complete as at the locality cited.

The lithology of the conglomerate can be well observed in a knoll on the northwest side of the Hope-Johnsonburg road, 1.2 miles northwest of Hope (Fig. 1, no. 4) and in a road-cut on the highway over Jenny Jump Mountain, 0.8 mile south of Hope (Fig. 1, no. 6). There is no trace of the conglomerate in the Jacksonburg exposures of the Hackettstown lowland 9 miles to the southeast, nor is it represented in the Jacksonburg type section 5 miles to the northwest. The size of the cobbles and the absence of these beds in nearby areas across the strike indicate the source of the material to be a local uplift, perhaps in the nature of a fault scarp. The conglomerate is provisionally assumed to be of Hull age, deposited in the advancing sea after the hiatus previously postulated from evidence at Jacksonburg and Portland. A similar conglomerate (Lacolle) from southern Quebec of Sherman Fall age has recently been described by T. H. Clarke and H. W. McGerrigle.[22]

JACKSONBURG-KITTATINNY RELATIONS

Four miles southwest of the conglomerate localities described above, the basal Jacksonburg is beautifully exposed in a large abandoned limestone quarry half a mile southwest of Sarepta (Fig. 1, no. 7). Black basal Jacksonburg beds overlie the light-colored magnesian beds of the Kittatinny terrane with angular unconformity (Pl. 1, fig. 2). The lowest

[21] H. B. Kümmel: *op. cit.*, p. 87-90.

[22] T. H. Clarke and H. W. McGerrigle: *Lacolle conglomerate: a new Ordovician formation in Southern Quebec*, Geol. Soc. Am., Bull., vol. 47 (1936) p. 665-674.

FIGURE 1. TOPOGRAPHIC EXPRESSION OF THE ORDOVICIAN FORMATIONS
Hills on extreme right underlain by Martinsburg shales, even-crested swell in center by Jacksonburg limestone, and fields on extreme left and in foreground by Beekmantown limestone.

FIGURE 2. QUARRY NEAR SAREPTA, NEW JERSEY
Dark beds of Jacksonburg formation (upper part of quarry face) overlie light-colored magnesium limestones of the Kittatinny. Lowest Jacksonburg beds contain large angular cobbles of the Kittatinny.

FIGURE 3. QUARRY IN SOUTHERN PART OF NORTHAMPTON, PENNSYLVANIA
Beekmantown limestone complexly folded.

ORDOVICIAN FORMATIONS

FIGURE 1. MARTINSBURG SHALE LYING UNCONFORMABLY ON THE JACKSONBURG
One mile north of Fogelsville, Pennsylvania. Two hammers are at the contact. Angular divergence of 15 degrees between Jacksonburg beds (right) and Martinsburg beds (left).

FIGURE 2. QUARRY OF THE LONE STAR CEMENT COMPANY
In west portion of Nazareth. Beds on right represent "cement limestone" facies of the Jacksonburg, those on the left "cement rock" facies. Boundary between the two descends diagonally to the left from notch in quarry skyline.

Jacksonburg beds consist of a coarse conglomerate of angular cobbles of magnesian limestone which attain a size of 2 feet in the lowest beds but become mere angular fragments in beds 5 feet above the base. These small fragments are not ubiquitous in the lower beds but are present at a horizon 40 feet above the base. *Leperditia* occurred at 38 feet. This conglomerate is not the same as that already described. Both the cobbles and the fragments are angular, showing little transportation, in contrast with the well-rounded cobbles of the intercalated conglomerate at Hope and Johnsonburg. It occurs in basal beds containing *Leperditia,* whereas the conglomerate at Hope overlies non-conglomeratic limestone beds, which are believed to be higher than the *Leperditia* zone.

The angular unconformity at the base of the Jacksonburg is much more pronounced at Sarepta than at any other locality known to the writer. The basal Jacksonburg beds lap eastward over the eroded edges of the Kittatinny limestone.

A belt of Jacksonburg limestone occurs in the center of the Phillipsburg-Washington limestone lowland (Fig. 2), the structure being essentially synclinal with the argillaceous limestones of the Jacksonburg preserved in the middle. The Jacksonburg was formerly quarried by the Edison Cement Company, 1¼ miles southwest of New Village and one mile east of Stewartsville (Fig. 2, no. 1). At the southeast end of the quarry the contact with the Beekmantown is exposed in four places, the beds being overturned so that the Beekmantown now appears on the top. The contact is a sharp break, but without angular relationship or basal conglomerate. The lowermost Jacksonburg beds are highly fossiliferous, although the fossils are mostly fragmental. The following forms were identified:

Sowerbyella sp.	f
Dinorthis pectinella (Emmons)	f
Dalmanella rogata (Sardeson)	c
Parastrophina hemiplicata (Hall)	f

The presence of *Parastrophina hemiplicata* in this lower Trenton faunule indicates that the *Leperditia* beds of the Jacksonburg type section are not here represented and that the lowest beds are probably of Hull age.

The fossiliferous beds are high-grade gray limestone, but these change to shaly black argillaceous limestone with an irregularly developed slaty cleavage 10 to 20 feet above the base of the formation. The thickness of the Jacksonburg as estimated from this and nearby localities is at least 200 feet. The Martinsburg shales are nowhere preserved, however, so that it is not possible to tell how much of the total Jacksonburg is represented by this figure. The change from the lower pure limestone to the higher argillaceous beds is gradational. No detailed section has been

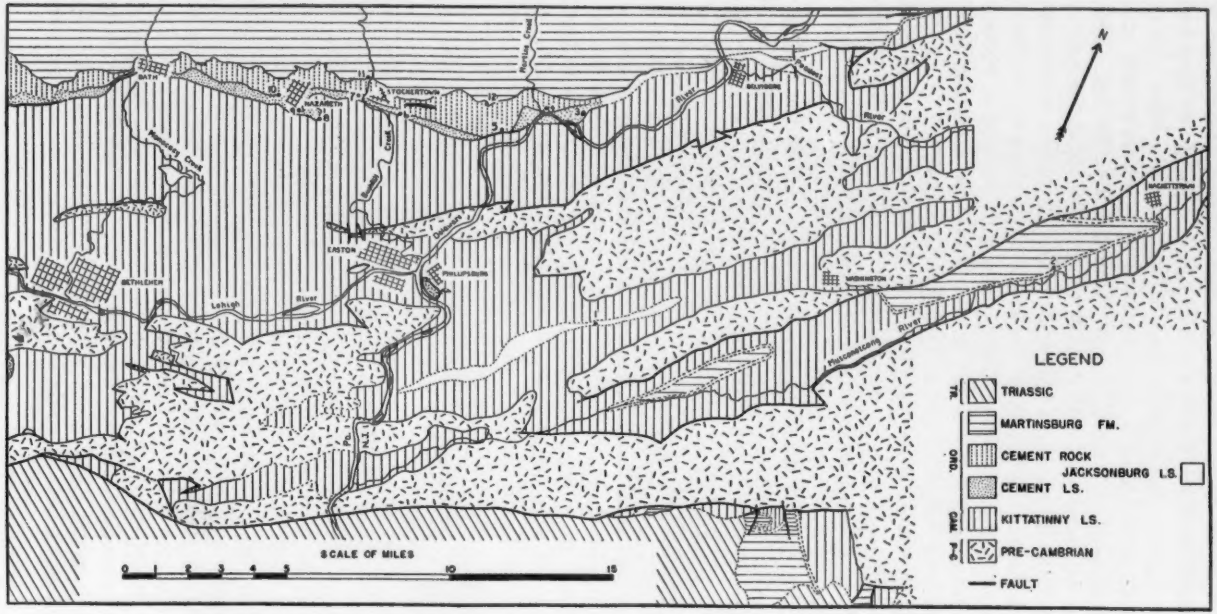

FIGURE 2.—*Geologic map of Jacksonburg belts in western New Jersey and eastern Pennsylvania*

Jacksonburg boundaries in New Jersey after H. B. Kümmel, in Pennsylvania by the writer. Other boundaries compiled from published maps by B. L. Miller, F. B. Peck, and state geologic maps of New Jersey and Pennsylvania. In Pennsylvania the cement-rock facies and the cement-limestone facies of the Jacksonburg formation are mapped separately.

measured in this quarry, as folds are visible in the west face, but elsewhere the bedding is concealed by slaty cleavage.

NORTHERN AND EASTERN LOCALITIES

In extreme northern New Jersey, Weller's Jacksonburg collections from the vicinity of Hamburg [23] show the presence of the Hull and Sherman Fall, but the lower or *Leperditia* beds are absent, as they are in the Edison Cement Company quarry. In the Green Pond Mountain region of north-central New Jersey, no Mohawkian limestones are present beneath the Silurian Green Pond conglomerate, but their absence is to be explained by the large pre-Silurian unconformity. Erosion progressed so far that the Green Pond conglomerate rests on the pre-Cambrian in much of the region.[24]

JACKSONBURG-MARTINSBURG RELATIONS

In all published accounts of the Jacksonburg both in New Jersey and Pennsylvania, it has been stated to pass upward into the Martinsburg shales without disconformity. A gradual increase in argillaceous content occurred with no sharp line of demarcation between the two formations. However, no continuous exposures from the Jacksonburg into the Martinsburg had been observed. One such exposure occurs in the Hackettstown limestone lowland, where there is a disconformity between the two formations and not a gradational change. This important exposure is on an abandoned loop of the Washington-to-Hackettstown highway (State Highway 24), 0.7 mile northeast of Stephensburg (Fig. 2, no. 2). Basal Martinsburg shales crop out along the old road for 100 yards back from the new highway, the beds being overturned so that they dip 15° NW. Near the end of the bluff, black argillaceous limestones of the Jacksonburg lie disconformably beneath the shales, although they appear on top, due to the overturning. The exact contact was dug out and proved to be a clay bed, 2 inches thick, separating the unweathered black limestone from the slightly weathered shales. The clay showed stratification parallel to the rock layers and was formed by solution of the impure limestone by water seeping along the contact. A suggestion of slippage is manifested by slightly contorted quartz stringers just below the contact, but this is expectable in rocks which have been folded to the extent of being overturned. The writer concluded that the slippage was minor and that this could not be considered a fault contact. Correspondence with Meredith Johnson, of the New Jersey Survey, established the fact that Kümmel had made observations on this exposure in his field

[23] Stuart Weller: *The Paleozoic faunas*, N. J. Geol. Surv., Rept. Paleont., vol. 3 (1903) p. 38.

[24] H. B. Kümmel and Stuart Weller: *The rocks of the Green Pond Mountain region*, N. J. Geol. Surv., Ann. Rept. (1901) p. 11.

notes subsequent to the publication of his Cement Report and the Raritan folio. He noted the overturning and that "the change from the cement rock to the shale or slate seems to be quite abrupt."

The presence of a hiatus between the Jacksonburg and the Martinsburg raises the question as to the amount of time represented thereby, and the age of the basal Martinsburg. The youngest fauna recognized in the Jacksonburg of New Jersey is of Sherman Fall age. The paucity of organic remains in the Martinsburg shales makes it difficult to establish the age of the basal beds. Weller [25] interpreted from his few faunules a Middle or Lower Ordovician age. Jacksonburg deposition continued well into Sherman Fall time, so that overlying Martinsburg cannot be older than later Sherman Fall. It is significant that the first deposition of shales (Canajoharie) in the Mohawk Valley of New York occurred in middle Sherman Fall time.[26] *Dalmanella edsoni* Bassler, which characterizes the lower Sherman Fall along Lake Champlain,[27] is present in limestone at the base of the Martinsburg shales in south-central Pennsylvania.[28]

RÉSUMÉ

At its type locality the Jacksonburg limestone of New Jersey consists of 135 feet of high-grade limestone, the lower 58 feet of which contain *Leperditia fabulites* Conrad and *Doleroides* sp. cf. *D. gibbosus* Billings. These beds may be of Black River age, as formerly interpreted, but seem more probably of Rockland age. The remaining 77 feet is of Hull and Sherman Fall age. A hiatus exists between the *Leperditia* zone and the upper division, which is marked by an abrupt faunal change at Jacksonburg and by a few small conglomeratic pebbles at Portland. This conglomerate is tentatively correlated with the thick coarse conglomerate in the Jacksonburg 5 miles southeast of the type locality. The *Leperditia* zone is absent at the Edison quarry in central New Jersey and probably throughout the southern Jacksonburg areas. According to Weller, it is also absent in extreme northeastern New Jersey. In both places, beds of probable Hull age rest on the Kittatinny limestone. In the southern areas of Jacksonburg the formation consists of high-grade limestone beds at the base, but the greatly thickened Sherman Fall beds are a black argillaceous limestone.

The Jacksonburg rests with great unconformity on the magnesian limestones of the Kittatinny terrane, usually containing angular cobbles or pebbles of the latter in the basal beds. It is overlain with some discon-

[25] Stuart Weller: *op. cit.*, p. 52.
[26] G. M. Kay: *Stratigraphy of the Trenton Group*, Geol. Soc. Am., Bull., vol. 48 (1937) p. 268.
[27] *Op. cit.*, p. 267.
[28] R. S. Bassler: *The Cambrian and Ordovician deposits of Maryland*, Md. Geol. Surv. (1919) p. 163, 165.

formity by the Martinsburg shales. Sufficient evidence is not available at present to delimit the margins of the *Leperditia* (Rockland ?) sea.

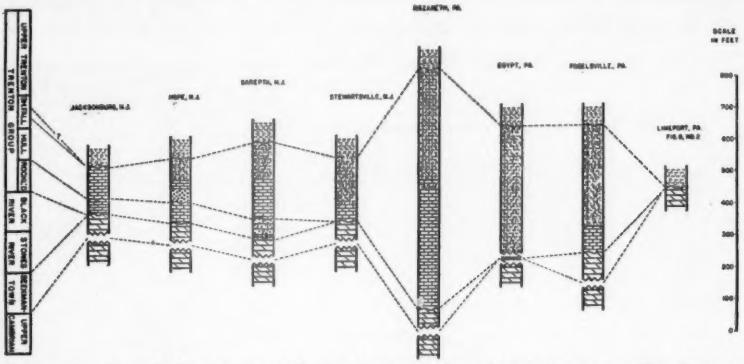

FIGURE 3.—*Correlation chart of Jacksonburg sections in northwestern New Jersey and eastern Pennsylvania*

Thicknesses for the lower Trenton limestones are approximate. For location of sections, *see* Figure 1 (Jacksonburg, No. 1; Hope; Sarepta, No. 7), Figure 2 (Stewartsville, No. 1; Nazareth), Figure 4 (Egypt; Fogelsville, No. 4; Limeport, No. 2).

The Hull and Sherman Fall seas extended over most of northern New Jersey. The tentative correlations for New Jersey sections are shown in Figure 3.

JACKSONBURG FORMATION IN PENNSYLVANIA

GENERAL DISTRIBUTION AND DESCRIPTION

The belt of Jacksonburg along the Paulins Kill-Jacoby Creek Lowland extends 4 miles into Pennsylvania (Fig. 1). Farther south the Sarepta belt of Jacksonburg (Fig. 2) crosses the Delaware River near Belvidere and becomes the main Jacksonburg belt of Pennsylvania. It extends continuously southwestward for 28 miles and has a much broader belt of outcrop for most of this distance than it does anywhere in New Jersey. This is due partly to an increased thickness and partly to more complex folding resulting in repetition of beds at the surface. Along this belt, 22 cement mills are located, several of which have not been operated for the past eight years. This belt is interrupted 4 miles west of the Lehigh River (Figure 4), and isolated areas of argillaceous limestone underlying the Martinsburg shales westward to the Susquehanna River have been designated as the Leesport limestone. The name Leesport was applied by Stose and Jonas,[29] its type section being at West Leesport on the Schuylkill River (Fig. 4, no. 1).

[29] G. W. Stose and A. I. Jonas: *Ordovician shale and associated lava in southeastern Pennsylvania*, Geol. Soc. Am., Bull., vol. 38 (1927) p. 505-536.

Intense deformation of the Jacksonburg has occurred in much of the Pennsylvania region. The complexity of folding in the argillaceous beds is commonly obscured by slaty cleavage, but a photograph of a quarry in the underlying, massive-bedded magnesian limestone shows the in-

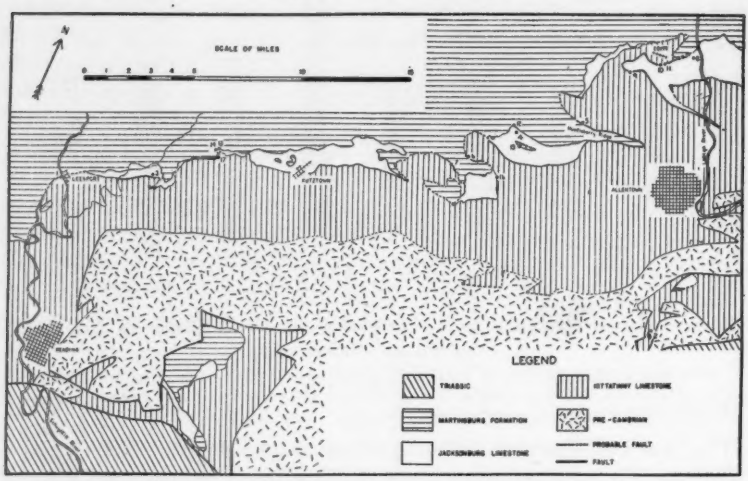

FIGURE 4.—*Geologic map of western Jacksonburg areas in Pennsylvania*

Boundaries of the Jacksonburg by the writer; all others from the state geologic map. The New Jersey term, Kittatinny, is here used to include dolomitic limestones of Cambrian and Beekmantown age.

tensity of the stresses to which the region has been subjected (Pl. 1, fig. 3). The deformation which produced the complex folding and faulting in all the Cambro-Ordovician sediments and the slaty cleavage in the argillaceous Jacksonburg limestone and the Martinsburg shales is believed by Behre [30] to be Taconic. Appalachian folding again deformed the Cambro-Ordovician strata, but the Appalachian structures were of much greater amplitude and lesser complexity.

AGE

Deformation has resulted in the distortion or complete obliteration of most of the fossils in the Jacksonburg of Pennsylvania. Several eastern localities have furnished recognizable faunules, but practically no fossils have been found west of the Lehigh River. The fossiliferous localities will be discussed in order from east to west along the main belt.

The first good collecting locality is 2 miles east of Martins Creek on the Belvidere Highway (Fig. 2, no. 3). Here, at a corner where the highway turns from west to north, an abandoned lane continues westward

[30] C. H. Behre, Jr.: *Slate in Pennsylvania*, Pa. Geol. Surv., 4th ser., Bull. M 16 (1933).

past a large barn at the corner and down a steep hill. A few ledges of crystalline Jacksonburg crop out in the lane on this hill slope from which the following forms were collected:

Pachydictya acuta Hall	a
Rhinidictya sp. cf. *R. mutabilis* Ulrich	a
Lingula elongata Hall	r
Lingula sp.	f
Sowerbyella sp.	c
Parastrophina hemiplicata (Hall)	c
Rafinesquina alternata (Conrad)	f
Strophomena sp.	r
Dalmanella rogata (Sardeson)	c
Pelecypod indet.	r
Isotelus fragment	r

These beds are the equivalent of the *Parastrophina* beds of the Jacksonburg type section which are of Hull age. They occur here near the base of the formation, as indicated by the low dips and the presence of a Beekmantown outcrop on the main highway near the corner.

West of this locality the argillaceous limestones of the upper part of the Jacksonburg are well exposed in the quarries of the Alpha Portland Cement Company (Fig. 2, no. 4) at Martins Creek and the Lehigh Portland Cement Company at Sandts Eddy (Fig. 2, no. 5). In the former, 350 feet (plus) of black fine argillaceous limestone averaging 66 per cent of $CaCO_3$ is exposed in the north quarry face. In the south quarry wall, *Prasopora orientalis* Ulrich and *Dalmanella rogata* (Sardeson) were found in beds of similar lithology but separated from the north wall rock by faults. The total thickness of the Jacksonburg near Martins Creek is 400 feet plus, this increase over the thickness in New Jersey being due principally to the divergence of the sediments and perhaps partly to the addition of younger Sherman Fall beds. On the other hand, the *Leperditia* beds of the type Jacksonburg seem to be lacking.

In the Sandts Eddy quarry of the Lehigh Cement Company the Beekmantown limestone is in fault contact with the argillaceous beds of the upper part of the Jacksonburg. From different parts of this quarry, the beds of which were not correlated due to folding and faulting, the following were collected:

Prasopora orientalis Ulrich	f
Sowerbyella sp.	f
Dalmanella rogata (Sardeson)	a
Dinorthis pectinella (Emmons)	r
Zygospira sp.	c
Brachiopod indet.	
Conularia sp.	r
Calymene sp.	r
Calliops sp. cf. *C. callicephalus* (Hall)	r
Trilobite indet.	

This faunule is composed principally of long-ranging species, but the frequent presence of *Prasopora orientalis* Ulrich indicates that it is referable to the Sherman Fall.

BASAL JACKSONBURG

More definite information as to the age of the basal Jacksonburg is obtainable 3 miles farther west. In the town of Churchville (Fig. 2, no. 6) a shallow abandoned quarry at the main crossroads exposes a few feet of crystalline limestone beds, the lowest of which carry small angular conglomeratic fragments of the underlying Beekmantown limestone. Five feet higher, *Receptaculites occidentalis* Salter was found. In the quarry of the Hercules Cement Company, one mile east of Stockertown (Fig. 2, no. 7), *Receptaculites* and *Dalmanella* sp. cf. *D. rogata* (Sardeson) were also found in beds near the base of the formation. The presence of *Receptaculites* in the basal Jacksonburg here, and of *Parastrophina* in near-basal beds east of Martins Creek, again shows the absence of the *Leperditia* zone of the type Jacksonburg.

This Hull-Beekmantown contact is exposed in the east end of the Trumbower Crushed Rock Quarry on the Tatamy road, one mile east of Nazareth (Fig. 2, no. 8). Unfortunately, the basal beds of the Jacksonburg are not fossiliferous. They rest on the Beekmantown disconformably, in one place containing angular pebbles and cobbles of the underlying magnesian limestone. Only a few feet away, however, no conglomerate is developed; the contact is so obscure that it is recognized with difficulty and so tight that a solid specimen may be collected across it. The absence of coarse basal conglomerates or interbedded conglomerates, such as occur in the New Jersey Jacksonburg, is a normal feature of the formation in Pennsylvania.

NAZARETH SECTION

One of the most complete sections of the Jacksonburg in the Lehigh Valley is exposed in the quarry of the Nazareth Cement Company in the eastern part of Nazareth (Fig. 2, no. 9). The lowermost beds of the Jacksonburg are not exposed, and an unknown thickness of the upper argillaceous limestones does not appear in the quarry. More than 600 feet of the formation is represented, however, before a fault is encountered. Distorted but recognizable fossils were found at three horizons.

The section measured here is as follows:

	Feet	Inches	Top of bed to base of section Feet
B5. Black, argillaceous limestone with slaty cleavage....	20	..	54
B4. Altered volcanic ash bed.....	...	2	34
B3. Black argillaceous limestone with slaty cleavage....	9	..	34

	Feet	Inches	Top of bed to base of section Feet
B2. Altered volcanic ash bed	...	5	25
B1. Black argillaceous limestone with slaty cleavage, probably higher than Bed A7	25	..	25
Fault			
A7. Black argillaceous limestone in massive beds with slaty cleavage; contains *Prasopora orientalis* Ulrich	150	..	610
A6. Black argillaceous limestone with slaty cleavage. Top layers contain *Prasopora* ? sp., *Dalmanella rogata* (Sardeson), *Dinorthis pectinella* (Emmons), *Parastrophina hemiplicata* (Hall), *Calliops callicephalus* (Hall), and *Rafinesquina* sp.	100	..	460
A5. Gray fine-grained, even-bedded limestone. Top layer contains *Mesotrypa* sp., *Dinorthis pectinella* (Emmons), *Parastrophina* sp., *Dalmanella rogata* (Sardeson), *Sowerbyella* sp., and *Rafinesquina* sp.	105	..	360
A4. Altered volcanic ash bed	...	13	255
A3. Gray fine-grained, even-bedded limestone	50	..	254
A2. Altered volcanic ash bed	...	1½	204
A1. Gray fine-grained, even-bedded limestone	204	..	204

No fossils were found in the lower 350 feet of the formation. The higher zones indicate the usual Sherman Fall, and possibly upper Hull, age for the argillaceous beds. The writer believes that the formation approximates this thickness in the region around the Hercules Company quarry, previously discussed, where *Receptaculites* was found in beds near the base, and that the entire 600 feet (plus) is probably of Hull and Sherman Fall age.

FACIES DIVISIONS

The Nazareth Cement Company quarry in the eastern part of Nazareth and the quarry of the Lone Star Cement Company in the western part of Nazareth (Fig. 2, no. 10) reveal most clearly the two facies of the Jacksonburg in the Lehigh Valley. The lithologic distinction is quite clear between the gray, massive-bedded (in places, crystalline) limestones of the lower part of the formation and the fine-grained, argillaceous, black limestones with slaty cleavage of the upper part of the Jacksonburg (Pl. 2, fig. 2). The high limestone beds commonly analyze from 75 to 95 percent $CaCO_3$, the argillaceous limestone beds from 60 to 70 percent of $CaCO_3$. This has led Wherry,[31] Miller,[32] and Peck [33] to divide the Jack-

[31] E. T. Wherry: *The early Paleozoic of the Lehigh Valley District, Pa.*, Science, n. s., vol. 30 (1909) p. 416.

[32] B. L. Miller: *Mineral resources of the Allentown Quadrangle, Pennsylvania*, Pa. Geol. Surv., 4th ser., Topog. Geol. Atlas no. 206, Allentown Quadrangle (1925).

[33] F. B. Peck: *Preliminary report on the talc and serpentine of Northampton County and the Portland cement materials of the Lehigh District*, Topog. Geol. Surv. Pa., Rept. 5 (1911).

sonburg into two formations. No name hitherto applied is in current usage, however, and the usual practice at present is to refer to the lower pure limestone as the "cement limestone" and the upper argillaceous limestone as the "cement rock".

The writer recognizes the value of distinguishing between the two facies, and has found it practicable to map them separately from Martins Creek westward to the Lehigh River, a distance of 20 miles. Weller and Kümmel also recognized the same lithologic divisions in the southern areas of Jacksonburg in New Jersey, but narrower belts of outcrop and few exposures within those belts did not make it feasible to map the two divisions.

The writer does not believe it advisable to give formational, or even member, status to these lithologic divisions for the following reasons: (1) At most localities, paleontologic evidence for establishing the time equivalence of the separate facies is lacking. (2) The incursion of clastic material was not contemporaneous throughout the whole area, so that the argillaceous facies at one locality does not represent the same interval as at another. The argillaceous facies clearly overlaps the limestone facies in a northwest direction in New Jersey and appears to do the same in eastern Pennsylvania. (3) Argillaceous beds occur in the upper part of the limestone facies, and pure limestone beds in the lower part of the argillaceous facies, so that the mapping of the boundary between the two is subject to considerable error except where continuous exposures cross the interbedded zone. The writer proposes, therefore, to retain the terms which are in common usage, and to refer to the "cement limestone facies" and "cement rock facies" of the Jacksonburg formation.

JACKSONBURG-MARTINSBURG RELATIONS

The Jacksonburg has been stated to grade upward into the Martinsburg without a definite break, but no contact between the two had been observed in Pennsylvania. It has been established that a disconformity exists between the two near Hackettstown, New Jersey. In Pennsylvania, two places have been found where the juxtaposition of the Martinsburg and the Jacksonburg is observable. In the bluff on the west bank of Bushkill Creek, 0.9 mile northwest of Stockertown (Fig. 2, no.11), the Martinsburg overlies the black argillaceous beds of the Jacksonburg with an angular divergence of 15 degrees. Slickensiding on the lower surface of the Martinsburg indicates movement of the shale in a northwest direction over the limestone. No conclusions are warranted from this exposure. It is of interest, however, because a short prospect tunnel driven in the limestone just beneath the contact exposes perfectly the slickensided undersurface of the Martinsburg.

One and a half miles west of Martins Creek, a small anticline exposes the Martinsburg-Jacksonburg contact, or near-contact, at three places (Fig. 2, no. 12). At creek level there appears to be angular divergence but a short distance up the bank the Martinsburg and Jacksonburg beds are parallel, although disconformity is distinct. The evidence here is contradictory; the hill slope is steep so that creep in the nearly solid Martinsburg beds may explain the observed angular divergence.

Observation at many other places where upper Jacksonburg crops out very near the foot of the shale hills has established the fact that the change from argillaceous limestone to non-calcareous shale is abrupt and not gradational. The highest Jacksonburg beds are highly calcareous and can be distinguished by eye, and by their effervescence with dilute acid, from the succeeding shale beds which are non-calcareous. A hiatus exists between Jacksonburg and Martinsburg beds; the Jacksonburg seas withdrew within Sherman Fall time. Uplift of Appalachia occurred so that, when the Martinsburg seas re-advanced, non-calcareous muds were deposited.

Stose and Jonas [84] have pictured and described an important exposure at Limeport (Fig. 4, no. 2) where Martinsburg shale rests with angular unconformity on Upper Cambrian limestone. The problem presented by the Limeport unconformity becomes one of determining whether the Jacksonburg was deposited and eroded out during the post-Jacksonburg hiatus, or whether the Jacksonburg seas did not cover this area. The further problem of why the Beekmantown is absent is probably explained in that, elsewhere, erosion occurred in pre-Jacksonburg time, as evidenced by the inclusion of magnesian limestone pebbles in basal Jacksonburg beds. This erosion was locally profound at Sarepta, New Jersey, and, at Limeport, it apparently was sufficiently profound to remove all the Beekmantown. On the other hand, the basal Martinsburg beds of the Lehigh Valley and of New Jersey are not conglomeratic, nor are the initial sediments coarse grained. It may be inferred, therefore, that the Martinsburg seas advanced over lands which were low lying and had not recently been subjected to deep erosion. It seems improbable that approximately 600 feet of Jacksonburg was removed in the pre-Martinsburg hiatus, and more likely that the Limeport area was a low-lying land during Jacksonburg time. All Chazyan time is available for the erosion of the Beekmantown. The writer believes that the Martinsburg lies on the Upper Cambrian at Limeport because the Beekmantown was eroded out in pre-Jacksonburg time and that this area was low-lying land in Jacksonburg time.

[84] G. W. Stose and A. I. Jonas: *Ordovician shale and associated lava in southeastern Pennsylvania*, Geol. Soc. Am., Bull., vol. 38 (1927) p. 518-520.

RELATIONS OF THE JACKSONBURG AND LEESPORT FORMATIONS

GENERAL STATEMENT

The Leesport formation has been designated by Stose and Jonas [35] to include the argillaceous limestones ("cement rock") which underlie the Martinsburg shales and overlie Beekmantown or Stones River limestones in the area between the Lehigh and the Susquehanna river. The relations of these beds to the Chambersburg limestone of south-central Pennsylvania and to the Jacksonburg limestone of eastern Pennsylvania are confused, and paleontologic evidence is lacking. Hence, Stose and Jonas deemed it inadvisable to imply any correlation by applying either of these names, and designated these beds as a new formation. They chose the exposure in the cut of the Reading Railroad at West Leesport on the Schuylkill River (Fig. 4, no. 1) as the type section.

Detailed work by the writer in the area between the Schuylkill and the Lehigh rivers has uncovered facts which throw some light on this problem. The results of the work have led to the conclusion that the argillaceous limestones east of the Schuylkill should all be referred to the Jacksonburg rather than to the Leesport. The reasons governing this decision will be discussed under the following headings: (1) Leesport type section; (2) Lithology of the Leesport "cement rock"; (3) Continuity of the "cement rock"; (4) Structure of the Egypt region; (5) Relation to overlying and underlying beds. A discussion of this problem will indicate other stratigraphic and structural relations of the eastern part of the Leesport belt, as the writer interprets them.

LEESPORT TYPE SECTION

The choice of the type section for the Leesport formation was unfortunate, in that the section is one of complex structure and the lithology is quite unlike the typical "cement rock" elsewhere in the Leesport belt. Figure 5 is a structure section traced from a panoramic photograph of the type section; the blank areas in the illustration represent covered areas in the railroad-cut. One hundred and thirty-four feet of beds is exposed, but the interrelationship of different parts of the exposure is obscure. The contact of the limestones with the Martinsburg shale is a fault contact, as shown by slickensides and pronounced drag of the thin-bedded limestones along the fault plane. Another fault of considerable proportions may exist between the lower thin-bedded limestones and the massive-bedded calcareous sandstones. At first glance the structure appears to be a syncline, but the thin-bedded limestones at the north end of the section have more prominent shale partings than those at the south end. Further, such resistant beds as the 35 feet of calcareous massive sandstone

[35] *Op. cit.*, p. 511.

should crop out if they were duplicated on the north side of the 70-foot series of limestones. Their present outcrop is limited at the top by the projected fault plane at the base of the Martinsburg. A fault, therefore, seems more logical than a syncline as an explanation of the change of dip, and it also accounts for the covered area. The nearest exposure of older

FIGURE 5.—*Type section of the Leesport formation*

Exposed in the cut of the Reading Railroad at West Leesport, Pennsylvania. Traced from a panoramic photograph.

rocks lies half a mile to the south, where Beekmantown is present in a large quarry.

Lithologically the rocks in this section are strikingly different from the "cement rock" which has been called Leesport in areas to the east. The platy limestones at the north end of the section closely resemble limestones interbedded in the Martinsburg.[36] The massive calcareous sandstones also have their lithologic counterpart in limestones which have provisionally been interpreted as interbedded within the Martinsburg formation. The 70 feet of thin-bedded limestones at the south end of the cut are the only rocks exposed which are lithologically similar to the "cement rock". The writer is uncertain as to the stratigraphic position of the units in this section. If the name Leesport is used for these enigmatic rocks, the "cement rock" to the east should be given another designation.

LITHOLOGY OF THE "CEMENT ROCK"

The "Leesport cement rock" in areas between the Lehigh and the Schuylkill rivers is identical with the cement rock facies of the Jacksonburg described earlier in this paper. The rock exposed in the quarry of the Allentown Cement Company at Evansville (Fig. 4, no. 3) or of the Lehigh Cement Company at Gogelsville (Fig. 4, no. 4) cannot be distinguished by eye or by chemical analysis from that utilized for cement by plants operating in the Jacksonburg belt. On the other hand, it is strikingly different from most of the rock exposed in the Leesport type section. In the Lehigh River-to-Schuylkill River region, where paleontologic evidence fails, other bases for designation of formations must be relied upon.

[36] R. L. Miller: *Martinsburg limestones in eastern Pennsylvania*, Geol. Soc. Am., Bull., vol. 48 (1937) p. 93-112.

Using lithology as a basis, Jacksonburg would appear to be a more applicable name than Leesport for the cement rock of this area.

CONTINUITY OF THE "CEMENT ROCK" BELT

One of the main reasons why it has not seemed feasible in the past to apply the name Jacksonburg to the western "cement rock" areas is because of the discontinuity of the belt of outcrop west of the Lehigh River. This discontinuity is much more striking on earlier geologic maps than on Figure 4 of this report. Detailed mapping has demonstrated that the belt of cement rock is continuous from Monterey to a point within one mile of the Schuylkill River. The width of the belt varies greatly, but it does not seem to be entirely absent anywhere along this 15-mile stretch.

Between Monterey and the Legih River it is absent in several places. However, at most of these localities, faults are known to occur. Huckleberry Ridge (Figure 4) seems to be a graben of Martinsburg in contact with the Beekmantown on both sides. A fairly deep drilling on the top of the ridge encountered only shale at depths well below the level of the limestones on either side.[37] A series of springs lies along the southern margin of the Martinsburg belt. A quarry, 150 yards from the northern margin of the shales (Fig. 4, no. 5), exposes magnesian limestones dipping 11 degrees in a direction N. 82° E. These relations seem better explained by parallel faults bounding the shale ridge than by a tightly compressed syncline. A quarry at the foot of the shale hills, 1.1 miles south of Haafsville (Fig. 4, no. 6), displays an unusual development of closely spaced joints and of quartz veins in dolomitic limestones, which is taken as evidence of a nearby fault. A recumbent fold in the west end of the quarry suggests that this is a low-angle thrust from the southwest. There is similar evidence for a fault on the north side of the Maxatawn area (Fig. 4, no. 7), which may be the same low-angle thrust. Outcrops are few near the foot of the shale hills in each of the areas. The writer has mapped cement rock only where he has definite evidence that it exists; it may, therefore, be somewhat more extensive than the map shows. Whether the faults are of sufficient magnitude to explain the partial or complete absence of the cement rock in these three areas is not certain, but it is quit possible.

Two other explanations or alternatives were considered to account for the absence of the cement rock. Non-deposition did not seem to apply satisfactorily, because the "cement rock" diminishes from a thickness of several hundred feet to zero in short distances, and because there are no facies changes such as might be expected if the "cement rock" seas were lapping up on numerous islands or arches. A disconformity is known to occur at the base of the Martinsburg, but if erosion had been sufficiently

[37] B. L. Miller: *Personal communication.*

profound to remove all the Jacksonburg a succeeding conglomerate might be expectable. Basal conglomerates have not been found in the Martinsburg in this region. The change from considerable thickness to zero in short distances is also more suggestive of faulting than of unconformity.

STRUCTURE OF THE EGYPT REGION

The region in the vicinity of Egypt (Fig. 4) forms the missing link in the establishment of the relations between the main Jacksonburg belt east of the Lehigh River and the "cement rock" areas west of the Lehigh, which have been called Leesport. The structure in this vicinity is extremely complex. Stose and Jonas [38] collected trilobites from massive siliceous limestones near the foot of the Martinsburg escarpment, which were identified by Ulrich as *Leiostegium* and referred to his Upper Ozarkian. This would be above the Upper Cambrian division of the Kittatinny terrane (called Allentown limestone in this region) and below the Beekmantown division.

The writer has collected *Prasopora* from the stripped area at the west end of the Whitehall Cement Company quarry at Cementon (Fig. 4, no. 8), which establish this as being the cement-rock facies of the Jacksonburg of middle Trenton (Sherman Fall) age. This belt of cement rock continues 4 miles west of the Lehigh River and does not differ from the cement rock east of the river. The entire area is bounded by faults, which are clearly visible in two places (Fig. 4, nos. 9 and 10). A thrust fault was temporarily well exposed in a ditch along the road ascending the hill on the south side of Coplay Creek, 0.4 mile south of Egypt (Fig. 4, no. 11). The west-facing escarpment of Martinsburg just north of Egypt was also mapped by Peck [39] as a possible fault.

The cement-rock beds at the foot of the Martinsburg hills west of Egypt seem identical with the cement-rock beds south of Egypt, from which they are separated by a half-mile gap. The evidence of abundant faulting in this region is so clear that it seems logical to interpret the Martinsburg as being in fault contact with the *Leiostegium* beds of Stose and Jonas, even though critical exposures which reveal the fault are lacking. The absence of Beekmantown limestone, which is present to the east, west, and south, also suggests major faults. According to this interpretation the Jacksonburg belt is interrupted in the Egypt region because of faulting and continues half a mile farther west in its normal position at the base of the Martinsburg shale hills.

[38] G. W. Stose and A. I. Jonas: *op. cit.*, p. 511-512.

[39] F. B. Peck: *Preliminary report on the talc and serpentine of Northampton County and the Portland cement materials of the Lehigh District*, Topog. Geol. Surv. Pa., Rept. 5 (1911).

RELATION OF THE LEESPORT "CEMENT ROCK" TO OVERLYING AND UNDERLYING FORMATIONS

The contact of the "Leesport cement rock" and the Martinsburg shales is exposed in a road-cut on the east side of Hassen Creek, one mile north of Fogelsville (Fig. 4, no. 12). Both the shales and the "cement rock" are thin-bedded and black, but the lithologic change is abrupt, as shown by the vigorous effervescence of the argillaceous limestone and the non-effervescence of the shale in specimens an inch apart on opposite sides of the contact. An angular divergence of 15 degrees is present, as shown in Figure 1 of Plate 2. There is a slight curving of the Jacksonburg beds beneath the shale at one place along the contact, which might be interpreted as drag on a fault but is believed to be minor slippage related to folding.

The Martinsburg-"Leesport cement rock" contact is again exposed in two places, 1.4 miles north of Moselem Springs (Fig. 4, nos. 13 and 14), where Moselem Creek enters the shale belt. One of these is clearly a fault and the other a probable fault, so that they offer no supplementary information. At many localities where "cement rock" crops out near, though not in actual contact with, the Martinsburg shales, no gradational change from the argillaceous limestone lithology to shale has been observed.

The evidence indicates that the cement rock between the Lehigh and the Schuylkill rivers unconformably underlies the Martinsburg as does the Jacksonburg cement rock east of the Lehigh River. In the area from Lebanon to Womelsdorf west of the Schuylkill River, Stose and Jonas note that the "Leesport limestone and Stones River limestone underlie the shale in apparent conformity."[40] This relationship makes it preferable to refer the eastern "Leesport cement rock" belt to the Jacksonburg.

The "cement rock"–Beekmantown contact is exposed in three places. In an abandoned quarry, half a mile south of the Fogelsville plant of the Lehigh Cement Company and 0.1 mile north of U. S. highway 22 (Fig. 4, no. 15), the lower beds are Beekmantown, the upper beds a crystalline, pure limestone of the lowest "cement rock" formation. The exact contact is so indistinct that it cannot be recognized in much of the quarry, but it was found in two places. The lowest bed of the Jacksonburg contains a few small angular pebbles of the underlying beds. Similar relations exist in another quarry along a northeast-southwest trending lane, 1.4 miles north-northwest of Trexlertown (Fig. 4, no. 16). In part of this quarry an interbedded zone, one foot thick, of attending layers of Beekmantown "cement rock" lithology lies just above the contact, indicating that part of the material of the lowest "cement rock" was ob-

[40] G. W. Stose and A. I. Jonas: *op. cit.*, p. 510.

tained by the reworking of the Beekmantown. In a new road-cut along the highway, 1.2 miles north of Moselem Springs (Fig. 4, no. 17), the contact is again exposed. The interbedded zone is 3 feet thick, but the outcrop supplies little further information, as the beds are much disturbed and there has been considerable flowage of the thin-bedded "cement rock" around the massive blocks of the Beekmantown. The evidence indicates that the "cement rock" formation overlies the Beekmantown limestone disconformably in this region.

LEESPORT BELT WEST OF THE SCHUYLKILL RIVER

Four reasons have been advanced in support of the view that the "cement rock" beds east of the Schuylkill River should be referred to the Jacksonburg formation. West of the river, other areas occur in which the cement rock lithology is prominent. The enigmatic Leesport type section intervenes, however, and other complicating factors are present. These make it inadvisable to include the western areas of "cement rock" in the Jacksonburg formation unless the relations are worked out in more detail and prove to be similar to those in the eastern region just described.

The probable correlation of sections of the Jacksonburg in Pennsylvania with each other and with the Jacksonburg of New Jersey is shown in Figure 3.

ALTERED VOLCANIC ASH BEDS

Beds of altered volcanic ash are present in the Jacksonburg at several localities in Pennsylvania, but only one has been observed in New Jersey. This is probably due to the greater abundance in Pennsylvania of large quarries which furnish extensive unweathered exposures of the Jacksonburg, rather than to any rarity of altered volcanic ash beds in New Jersey. The clay beds are present in rocks of both Hull and Sherman Fall age. Similar beds have been described from rocks of probable Sherman Fall age in central Pennsylvania by Whitcomb [41] and in Virginia by Rosencrans.[42] Others have been described from both Hull and Sherman Fall rocks in widely scattered localities of eastern United States by Kay.[43] Four different clay beds were noted in the Nazareth Cement Company quarry, two in the cement-limestone facies of the Jacksonburg, and two in the cement-rock facies. The thickest one is a 13-inch bed within the cement-limestone facies, which is here probably of Hull age. Commonly, the clay beds are only one or two inches thick.

Attempts to correlate the altered volcanic ash beds in different locali-

[41] Lawrence Whitcomb: *Correlation by Ordovician bentonite*, Jour. Geol., vol. 40 (1932) p. 522-534.

[42] R. R. Rosencrans: *Bentonite in Northern Virginia*, Wash. Acad. Sci., Jour., vol. 23 (1933) p. 413-419.

[43] G. M. Kay: *Distribution of Ordovician altered volcanic materials and related clays*, Geol. Soc. Am., Bull., vol. 46 (1935) p. 225-244.

ties have not been successful because the thickness and spacing of the clays seem to vary over short distances and because paleontologic evidence, by which tentative correlations can be checked, is lacking. The following chart shows the localities at which altered volcanic ash beds have been observed:

Locality	Thickness in inches	Probable age of enclosing rocks
1. Vulcanite Cement Co. quarry, Alpha, N. J.		
Bed 1—NW corner of quarry	.5	Sherman Fall
Bed 2—14 feet above Bed 1	3.	Sherman Fall
2. Lehigh Cement Co. quarry, Sandts Eddy, Pa. Bed in north end of quarry	.25	Sherman Fall
3. Hercules Cement Co. quarry, Stockertown, Pa. West face	.25	Sherman Fall
4. Nazareth Cement Co. quarry, Nazareth, Pa.		
Bed 1—155 feet below top of cement limestone facies	1.5	Hull
Bed 2—105 feet below top of cement limestone facies	13.	Hull
Bed 3—north face of quarry in cement rock facies. Separated from Beds 1 and 2 by a fault	5.	Sherman Fall
Bed 4—9 feet above Bed 3	2.	Sherman Fall
5. Lone Star Cement Co. quarry, Nazareth, Pa. East face. 110 feet below top of cement limestone facies. Same as Bed 2 of Nazareth quarry	1.5	Hull
6. Keystone Cement Co. quarry, Bath, Pa. Southwest corner. 25 feet below top of cement limestone facies.	1.5	Hull?
7. Bath Cement Co. quarry, Bath, Pa.		
Bed 1—south face of south quarry in cement limestone facies. Same as Bed 2 in Nazareth Co. quarry?	13.	Hull
Bed 2—30 feet above Bed 1	1.	Hull
8. Universal Atlas Cement Co., Northampton, Pa. Southeast corner of quarry north of highway. Questionable ash bed	1.	Sherman Fall
9. Lehigh Cement Co. quarry, Fogelsville, Pa. Top of face in northeast corner of quarry.		
Bed 1	1.	?
Bed 2—questionable. 8 feet above Bed 1	.5	?
10. Abandoned cement rock quarry near Schlemville, Pa.	3.	?

SUMMARY AND CONCLUSIONS

In northwestern New Jersey the Jacksonburg formation consists, in its upper part, of beds of Hull and Sherman Fall age. These are high-calcium limestones in the northern localities but change to argillaceous limestones and thicken southward. They are disconformably overlain by the Martinsburg shales, the basal beds of which thus cannot be older

than middle Sherman Fall in this region. A disconformity is present beneath the upper beds of the Jacksonburg and is succeeded locally by an indistinct conglomerate; a thick, coarse conglomerate in other localities is believed equivalent. Beneath the disconformity in northern and western localities are *Leperditia*-bearing beds. The age of these is uncertain, but they are provisionally referred to the Rockland. The Jacksonburg unconformably overlies the magnesian limestones of the Kittatinny terrane.

In Pennsylvania, faunas of Hull and Sherman Fall age are present. In the main belt of the Jacksonburg, the *Leperditia* zone has not been recognized, and near Stockertown the basal beds seem to be Hull. The formation has been divided into two parts on the basis of lithology, the lower being the "cement limestone" facies and the upper the "cement rock" facies. These divisions are not given formational or member status, for their limits have not been accurately established, and the lithologic boundary probably transects time lines. The "cement rock" east of the Schuylkill River, which has been called Leesport, is assigned to the Jacksonburg. Volcanic activity during the lower Trenton is evidenced by the presence in the Jacksonburg of beds of altered volcanic ash.

A disconformity and boulder conglomerate within the Jacksonburg and an unconformity at the base of the Martinsburg indicate movement of lands and seas in lower Trenton time. Gradational overlap of the argillaceous limestone facies on the pure limestone facies from south to north is evidenced in New Jersey. This change is accompanied by convergence in the same direction; the convergence is gradual along the main belt from Pennsylvania into New Jersey, because of its more easterly trend. The incursion of muds is contemporaneous with a similar change of facies in the Mohawk Valley of New York and approximately synchronous with the introduction of clastics in Maryland and Virginia. An extensive uplift of land to the southeast in middle Trenton time is suggested, rather than a series of unrelated local uplifts. Investigations in southern New York and west of the Schuylkill River in Pennsylvania should furnish additional information, allowing better integration of the Trenton history of the eastern States.

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